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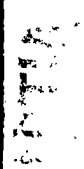
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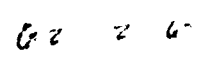
A REPRESENTATION OF THE TERRESTRIAL ATMOSPHERE FROM 100 KM TO 3000 KM

Marcel Nicolet

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SCIENTIFIC REPORT

on

"A REPRESENTATION OF THE TERRESTRIAL ATMOSPHERE FROM
100 KM to 3000 KM"

by

MARCEL NICOLET

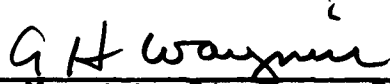
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IONOSPHERE RESEARCH LABORATORY

Approved for Distribution


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Abstract

The combined evidence from observational data and theoretical analysis are used to give a representation of the terrestrial atmosphere from 100 to 3000 km.

Quantitative information on the density of the upper atmosphere up to 1500 km, obtained from satellite data, shows that several constituents must control the vertical distribution in the heterosphere. The increase of the scale height with altitude corresponds to a non-linear gradient and is associated with the change of the atmospheric composition due to diffusion. In spite of being a rare constituent in the thermosphere, helium exercises a profound influence on atmospheric conditions above the thermopause. The possibility that hydrogen plays an important role below 2000 km cannot be accepted since its diffusion equilibrium cannot begin below the thermopause even if its concentration at 100 km is of the order of 10^7 cm^{-3} .

Molecular nitrogen is an essential constituent of the thermosphere since its vertical distribution corresponds to a diffusion equilibrium not too different from the air mixing distribution. Due to a recombination process, atomic nitrogen is subject, in the F1 layer, to strong diurnal variations in which a diffusion process is also involved.

Even allowing for the fact that dynamical equilibrium of oxygen is difficult to study, the observational evidence, and its theoretical interpretation, are now conclusive to the

effect that diffusion must be considered as an essential process counterbalancing dissociation processes.

The thermal conductivity of the heterosphere is such that the temperature varies rapidly from the exosphere down to an altitude below 200 km when solar conditions vary. All the evidence presented indicates that the variation of the temperature explains the atmospheric behavior on a world-wide scale.

A theoretically plausible atmosphere-model is presented in order to give a consistent picture for the atmospheric parameters.

1. Introduction

From the publication of atmospheric data as adopted by the Rocket Panel (see Whipple, 1954) to the publication of the COSPAR international reference atmosphere (see CIRA, 1961), various atmospheric models have been considered. Almost all models deal with data referring only to density (or pressure) and temperature. When a variation of the mean molecular mass is introduced, it generally corresponds to an arbitrary decrease with height and, therefore, a consistent vertical distribution of the composition cannot be obtained. Thus, it is necessary to distinguish, in the various published models, between the homosphere where the mean molecular mass can be taken as a constant and the heterosphere where the mean molecular mass varies with altitude according to laws subject to dissociation and diffusion. It can be shown that dissociation and diffusion processes are not important enough to change the mean molecular mass in the troposphere, stratosphere and mesosphere and these three atmospheric regions (the homosphere) can be considered together in the analysis of the atmospheric behavior. Above the mesopause, the effect of the dissociation of molecular oxygen must be introduced and above 100 km the effect of the diffusion of each atmospheric constituent requires a careful study.

In order to prepare a standard atmosphere or an atmospheric model the following equations must be used:

(1) the perfect gas obeys the law

$$p = n k T \quad (1)$$

in which p denotes the pressure, n the total concentration

or number density and T the temperature. If p is expressed in dyne cm^{-2} , n in cm^{-3} , T in $^{\circ}\text{K}$, k is the Boltzmann constant equal to 1.38×10^{-16} erg deg^{-1} .

(ii) the height-distribution of the pressure is governed by the statical equation

$$\frac{dp}{dr} = -g\rho \quad (2)$$

where r denotes the radius of the sphere at the altitude z above the earth's surface, ρ the density and g the acceleration of gravity.

The relation between ρ and n

$$\rho = n m \quad (3)$$

where m is the mean molecular mass leads to a general law, using (1) and (2),

$$\frac{dp}{p} = \frac{dn}{n} + \frac{dT}{T} = - \frac{dr}{kT/mg} \quad (4)$$

where $kT/mg = H$ is a quantity, having the dimensions of a length, called by Chapman the (local) atmospheric scale-height.

The equation must be used in the whole atmosphere where: (i) the effect of axial rotation of the planet is negligible, (ii) an escape of a constituent does not play an important role and (iii) the magnetic field cannot modify the vertical distribution.

For example, the effect of axial rotation is introduced at large distance r in an isothermal atmosphere for a constituent of concentration n_1 by the following equation

$$\frac{1}{n_1} \frac{\partial n_1}{\partial r} = - \frac{m_1 g}{kT} + \frac{m_1 \omega_e^2 r \sin^2 \theta_e}{kT} \quad (5)$$

in which ω_e is the angular velocity of the earth = 7.292×10^{-5} rad sec $^{-1}$ and θ_e is the colatitude.

The effect of an important escape in the exosphere requires a special analysis in which all trajectories of particles must be introduced after taking into account special processes such as ionization and charge transfers. Furthermore, in the thermosphere, the maximum flow which can be supported by diffusion must be determined. In the same way, the vertical distribution of charged particles at thermal energy in the exosphere requires an analysis of conditions affecting the distribution following the magnetic lines of force, such as the collisions leading to charge transfers, while the conditions of diffusion must be related to the effect of resonance charge transfers.

2. Published Atmospheric Models

It is not possible to describe all standard atmospheres which were proposed or even adopted since 1920. The background on the problem of standard atmospheres can be found in the introduction of 'The ARDC (Air Force Research and Development Command) Model Atmosphere' by Minzner and Ripley (1956) and will not be discussed here. Before procurement of the knowledge resulting from new data obtained from rocket-borne experiments, several attempts were made to extend to high altitudes the standard atmospheres which were applied to the low atmosphere. But, it must be pointed out that important modifications were made to the preceding conceptions when it became necessary to follow the vertical distribution of density

deduced from the rate of change of the orbital period of satellites. For example, if we consider the differences between the 1956 (Minzner and Ripley, 1956) and 1959 (Minzner, Champion and Pond, 1959) ARDC model atmosphere, the density at 600 km is 20 times greater in the 1959 model than in the 1956 model. However, the mean molecular mass was adjusted in such a way that it is about $M = 27.1$ at 180 km and decreases to 17 at 700 km. Such a vertical distribution of the mean molecular mass leads to an almost isothermal layer with $T = 1415 \pm 10^\circ\text{K}$ between 200 km and 300 km and with a continuous increase of the temperature gradient from 300 km up to 700 km; $\Delta T (700-600 \text{ km}) = 121^\circ\text{K}$.

The Russian model published by Mikhnevich et al (1959) from 225 km to 500 km also leads to an increase of the temperature gradient with the altitude. It is of the order of 1°K km^{-1} at 250 km and reaches 7°K km^{-1} at 700 km. It has been shown by Nicolet (1960 c) that such a vertical distribution of temperature in an atmospheric model cannot be accepted since it requires a downward heat flow by conduction which should be several times greater at 700-500 km than at 250-200 km. In fact, it was concluded by Nicolet (1960 b) that the time of conduction at very high altitudes is so short that there is a strong tendency to a vertical isothermy for the whole atmosphere. Various results obtained for isothermal atmospheres at various temperatures (Nicolet 1960 c, 1961 a) show that the observed vertical distribution of the atmospheric density can be followed above 200 km by atmospheric

models in which the temperature is practically constant. Furthermore, the diurnal variation of the density which was deduced from satellite observations by Jacchia (1959), Wyatt (1959), and Priester and Martin (1959) was explained (Nicolet 1961 a) by a variation of temperature between day and night, i.e. as a result of ultraviolet heating and a rapid cooling by conduction. Finally, anomalous behavior such as discontinuities in the vertical distribution of scale heights, densities and temperature were rejected since they result from the methods of analysis of observational data.

The COSPAR international reference atmosphere (CIRA, 1961) includes references to various proposals made in 1961. It is a matter of fact that there is now a tendency among the various authors to adopt the conditions of isothermy, but the mean molecular masses are arbitrarily assigned. Thus, it is not possible to deduce consistent data for the concentrations, and the only acceptable results above 100 km are the values of the density; which is the only parameter deduced from rocket and satellite observations. However, even when the densities for the proposed models are compared at certain heights, they are not consistent. Such inconsistencies, which generally occur near 200 km, do not allow a determination of physical parameters such as temperature and composition.

Finally, the high densities derived from the rate of change of period of the Echo satellite orbit cannot be explained by an extrapolation from the data at 800 km of the CIRA model. Atomic hydrogen alone cannot explain the observed

densities but a helium layer resolves the difficulty (Nicolet 1961 b).

3. Composition of the Homosphere

Analysis of the air at ground level and at different heights in the troposphere leads to a sufficient knowledge of the mixed composition of the homosphere. Table I gives the composition according to the best data available (see Nicolet, 1962).

Since molecular nitrogen is the principal constituent of the homosphere, and since its mass $M = 28$ is not too different from the mean molecular mass of the air $M = 29$, it is useful to consider the ratio of the concentrations of a constituent and molecular nitrogen. This ratio is given in Table I. When diffusion begins, there is a relative increase of hydrogen, helium, neon and atomic oxygen, and there is a relative decrease of argon, krypton and xenon. Carbon dioxide and methane are dissociated in the diffusion region. Hence, when the relative ratio of the hydrogen, helium, atomic oxygen concentrations and nitrogen concentration increases with height, calculation shows that the mass density of light constituents can exceed the nitrogen mass density at correspondingly higher levels according to the following order: atomic oxygen, helium and hydrogen. This will naturally affect the mean molecular mass of the heterosphere from $M = 28$ to the limit $M = 1$.

TABLE I
Composition of the Homosphere

Constituent	M	Air = 1	Nitrogen = 1	Isotope ratio
Molecular nitrogen	28	0.7808	1	$N^{14}/N^{15} = 273$
Molecular oxygen	32	0.2095	2.68×10^{-1}	$O^{16}/O^{18} = 490$
Argon	40	9.34×10^{-3}	1.20×10^{-2}	
Carbon dioxide (1)	44	3×10^{-4}	3.84×10^{-4}	$C^{12}/C^{13} = 89$
Neon (2)	20	1.82×10^{-5}	2.33×10^{-5}	$Ne^{20}/Ne^{21} = 354$
Helium (3)	4	5.24×10^{-6}	6.71×10^{-6}	$He^4/He^3 = 10^6$
Krypton (4)	84	1.14×10^{-6}	1.46×10^{-6}	
Xenon (4)	131	8.7×10^{-8}	1.11×10^{-7}	
Methane	16	1.5×10^{-6}	1.92×10^{-6}	
Molecular hydrogen (5)	2	6.0×10^{-7}	7.68×10^{-7}	

(1) C.D. Keeling, Tellus, 12, 200, 1960

(2) E. Glueckauf, Proc. Roy. Soc. A 185, 98, 1946

(3) E. Glueckauf and F.A. Paneth, Proc. Roy. Soc. A 185, 69, 1946

(4) E. Glueckauf and G.P. Kitt, Proc. Roy. Soc. A 234, 557, 1956

(5) E. Glueckauf and G.P. Kitt, Quarterly J.R.M.S., 83, 522, 1957

4. Physical and Chemical Transformations in the Heterosphere

The difficulty attached to calculations of physical parameters at different heights in the heterosphere comes from the uncertainty of the mean molecular mass effected by diffusion. This uncertainty is also increased because of the poor knowledge of the temperature distribution above 100 km. There is, further, the difficulty of the calculation of the atomic and molecular oxygen distribution which is the chief transformation of the lower thermosphere. The dissociation of molecular nitrogen may complicate the already difficult problem of the exact constitution of the heterosphere.

The theoretical study by Nicolet and Mange (1954) of the departure from photochemical equilibrium conditions led them to the recognition that, above the dissociation region, oxygen is not completely dissociated but still exists in molecular form. There is no doubt that this is because the major effect is that of diffusion which counterbalances the photodissociation (Nicolet, 1954, Mange, 1955). The rocket measurements by Byram, Chubb and Friedman (1957) show that about $5 \times 10^9 \text{ O}_2 \text{ molecules cm}^{-3}$ are present at 150 km, i.e. a density not less than 10% of the total density. The recent values obtained by Detwiler, Purcell and Tousey (1961) on the solar radiation between 2400-1215 Å lead to a dissociation rate coefficient of molecular oxygen $J_{\infty}(\text{O}_2)$ at zero optical depth of

$$J_{\infty}(\text{O}_2) = 4 \times 10^{-6} \text{ sec}^{-1} \quad (6)$$

From the photodissociation rate given by (6), it can be concluded that any concentration of O_2 greater than the photochemical equilibrium value requires about 2×10^5 sec in order to be reduced by 50 percent. Since the time of diffusion is shorter than the time of dissociation it is, therefore, clear that a diffusion distribution can be assumed in a first approximation for the vertical distribution of molecular oxygen above the dissociation region.

It is certain, however, that in an atmosphere where there is a large diurnal variation of the temperature that the vertical distribution of O_2 cannot be in a permanent distribution of diffusion. Thus, it can be said that some deficiency of molecular oxygen may exist during the daytime.

Other points of view have been adopted by several authors, who consider that the molecular oxygen concentration is very small above the F1 layer. For example, Danilov (1961 a) considers that the reactions which are important above the F1 peak are dissociative recombination



with a rate coefficient $\alpha_{O_2} = 10^{-6} \text{ cm}^3 \text{ sec}^{-1}$, and radiative association



with a rate coefficient $k_r = 5 \times 10^{-11} \text{ cm}^3 \text{ sec}^{-1}$. It is apparent that the magnitude of k_r is much too great for a radiative association (Bates 1961).

If (7) and (8) were to lead to the equilibrium equation

$$\frac{n^+(O_2)}{n^+(O)} = \frac{k_r n(O)}{\alpha_{O_2} n_e} \quad (9)$$

atomic oxygen ions should disappear very rapidly according to

$$-\frac{dn^+(O)}{dt} \approx k_r n(O) n^+(O) \quad (10)$$

It is clear from (10) that the ion concentration would be reduced to 1 percent of its sunset value in a very short time, i.e. a time given by

$$\tau \left[\frac{1}{100} n^+(O) \right] = 10^{11}/n(O) \text{ sec} \quad (11)$$

For $n(O) \approx 10^7 \text{ cm}^{-3}$, a time shorter than 10^4 sec would reduce the sunset concentration to less than 1 percent, i.e. no nighttime ionosphere at 350 km.

The extent to which molecular oxygen is dissociated depends on the importance of the loss process



in addition to the normal photodissociation (6) compared with the vertical transport, w , by diffusion. A continuity equation such as

$$\frac{dn(O_2)}{dt} = \frac{d[n(O_2)w]}{dt} - n(O_2) [J_\infty(O_2) + \gamma_{O_2} n^+(O)] \quad (13)$$

must be used in order to find the vertical distribution of O_2 . Since we maintain that rate coefficients of atom-ion interchange have the orders of magnitude given by Bates and Nicolet (1961), the diffusion distribution of molecular oxygen represents, in a first approximation, the vertical distribution

of that molecule. A diffusion distribution leads to a ratio 2 for the O_2 concentrations at 200 km and 250 km, while formula (9) leads to a ratio 25.

The dissociation of nitrogen is related, according to Nicolet (1960), to the vertical distribution of molecular oxygen. Without the presence of molecular oxygen, atomic nitrogen produced by dissociative recombination would not recombine. In fact, atomic nitrogen reacts with O_2



The rate coefficient b_1 of this reaction can be written as follows

$$b_1 = 1.5 \times 10^{-11} T^{1/2} e^{-3000/T} \quad (15)$$

showing that (14) depends strongly on the temperature, and is an important loss process of atomic nitrogen in the F1 region where the temperature is high and where there is still a sufficient number of O_2 molecules. Furthermore, nitric oxide, which is produced by (14), reacts very rapidly with another nitrogen atom as follows



Since such a reaction has practically no activation energy, its rate coefficient b_2 is written

$$b_2 = 1.5 \times 10^{-12} T^{1/2} \quad (17)$$

In chemical equilibrium, the following equation can be written

$$\frac{n(NO)}{n(O_2)} = 10^{-1} e^{-3000/T} \quad (18)$$

and the minimum life time $\tau(N)$ of a nitrogen atom (to reduce its initial concentration to 50%) is

$$\tau(N) = 5 \times 10^{10} / T^{1/2} e^{-3000/T} n(O_2) \quad (19)$$

Expression (19) shows that atomic nitrogen has a short life-time near the F1 peak leading to an important variation between day and night.

Different points of view have been adopted by various authors. For example, Hertzberg (1961), considers that there is a dissociation of molecular nitrogen in the F2 region while Danilov (1961 b) introduces the radiative association



with a rate coefficient of the order of $3 \times 10^{-11} \text{ cm}^3 \text{ sec}^{-1}$ with a production rate $\gamma(N) \approx 10^4 \text{ cm}^{-3} \text{ sec}^{-1}$. Again, such high production rates of N_2 and N are impossible, since the life time of an electron would be too short during dark hours.

Formula (11) can be applied to determine the minimum life-time of electrons. Thus, Equations (14) and (16) must be considered as representing the fundamental processes for neutral nitrogen. As far as atomic nitrogen ions are concerned, they are related to the photoionization $N_2 \rightarrow N^+ + N$ for $\lambda < 500\text{\AA}$.

5. Atomic Nitrogen and Nitric Oxide

Above the E layer where there is a rapid increase of temperature with height, the ionization of N_2 by ultra-violet radiation and the transformation by an ion-atom interchange of O^+ into NO^+ lead to the production of atomic nitrogen after dissociative recombination of the molecular ions

(see discussion by Nicolet, 1960). Photoequilibrium conditions lead to

$$n(N_2) I_{N_2} + Y n(O) I_0 = b_2 n(N) n(NO) \quad (21)$$

and

$$n(NO) = \frac{b_1}{b_2} n(O_2) \quad (22)$$

Since the ionization of atomic oxygen is not more important than the ionization of nitrogen, an idea of the conditions is obtained without considering the effect of atomic oxygen. An application based on (21) and (22) leads to the values of concentrations shown in Table II. The purpose of this Table is to show that the concentrations of N and NO in the thermosphere are small compared with the concentrations of N_2 and O_2 .

The life time of a nitrogen atom after sunset is short. Between 130 km and 220 km it is, according to (19), less than 10^3 sec, and there will be a downward transport of nitrogen atoms during the night from the upper F2 region to the F1 region. Thus, atomic nitrogen will play no practical role in the determination of the mean molecular mass of the thermosphere even if the rate of production is increased. A special study is required in order to investigate its vertical distribution which depends on varying boundary conditions at the bottom of the F region. It should be pointed out, however, that the atomic nitrogen concentration of Table II will vary according to the variation of the ultraviolet photons related to that of solar activity. As far as nitric oxide

Table IIAtomic Nitrogen and Nitric Oxide in Photoequilibrium

Altitude (km)	T (°K)	$n(N_2)$	$n(O_2)$	$n(NO)$	$n(N)$
130	447	1.8×10^{11}	3.3×10^{10}	4.0×10^6	3.3×10^6
140	570	7.7×10^{10}	1.3×10^{10}	6.6×10^6	3.5×10^6
150	692	3.8×10^{10}	5.9×10^9	7.7×10^6	3.5×10^6
160	813	2.1×10^{10}	3.1×10^9	7.8×10^6	3.2×10^6
170	907	1.3×10^{10}	1.8×10^9	6.7×10^6	3.2×10^6
180	980	8.9×10^9	1.1×10^9	5.4×10^6	3.2×10^6
190	1038	5.9×10^9	7.7×10^8	4.3×10^6	3.2×10^6
200	1081	4.2×10^9	5.3×10^8	3.3×10^6	3.3×10^6
220	1142	2.3×10^9	2.7×10^8	1.9×10^6	3.5×10^6
240	1182	1.3×10^9	1.4×10^8	1.1×10^6	3.7×10^6
260	1206	7.7×10^8	7.7×10^7	6.4×10^5	4.0×10^6
280	1222	4.6×10^8	4.3×10^7	3.7×10^5	4.3×10^6
300	1232	2.8×10^8	2.4×10^7	2.1×10^5	4.6×10^6

is concerned, its vertical distribution can be adapted to a diffusion distribution if the boundary conditions are chosen near the peak; in the neighborhood of 150 km.

6. Atomic Oxygen

Because the times involved in the recombinations of atomic oxygen increase rapidly with height above the mesopause, photoequilibrium cannot be reached. The altitude of the peak of atomic oxygen is not determined by photochemical conditions but by the effect of transport depending on

atmospheric conditions. (see paragraph 2.4 in Nicolet (1960 b)).

If we consider the following equation

$$n(O_2) J_2 = a n(M) n^2(O), \quad (23)$$

in which $n(O_2) J_2$ represents the production of oxygen atoms and $a n(M) n^2(O)$ their recombination by three-body collisions involving a third particle with concentration $n(M)$, we must find where photoequilibrium conditions can exist. The values of the rate coefficient J_2 are given in Table III when O_2 is subject to preliminary absorption by columns of 5×10^{16} to $5 \times 10^{20} O_2$ molecules cm^{-2} . If we adopt the usual coefficients, the vertical distribution of atomic oxygen for

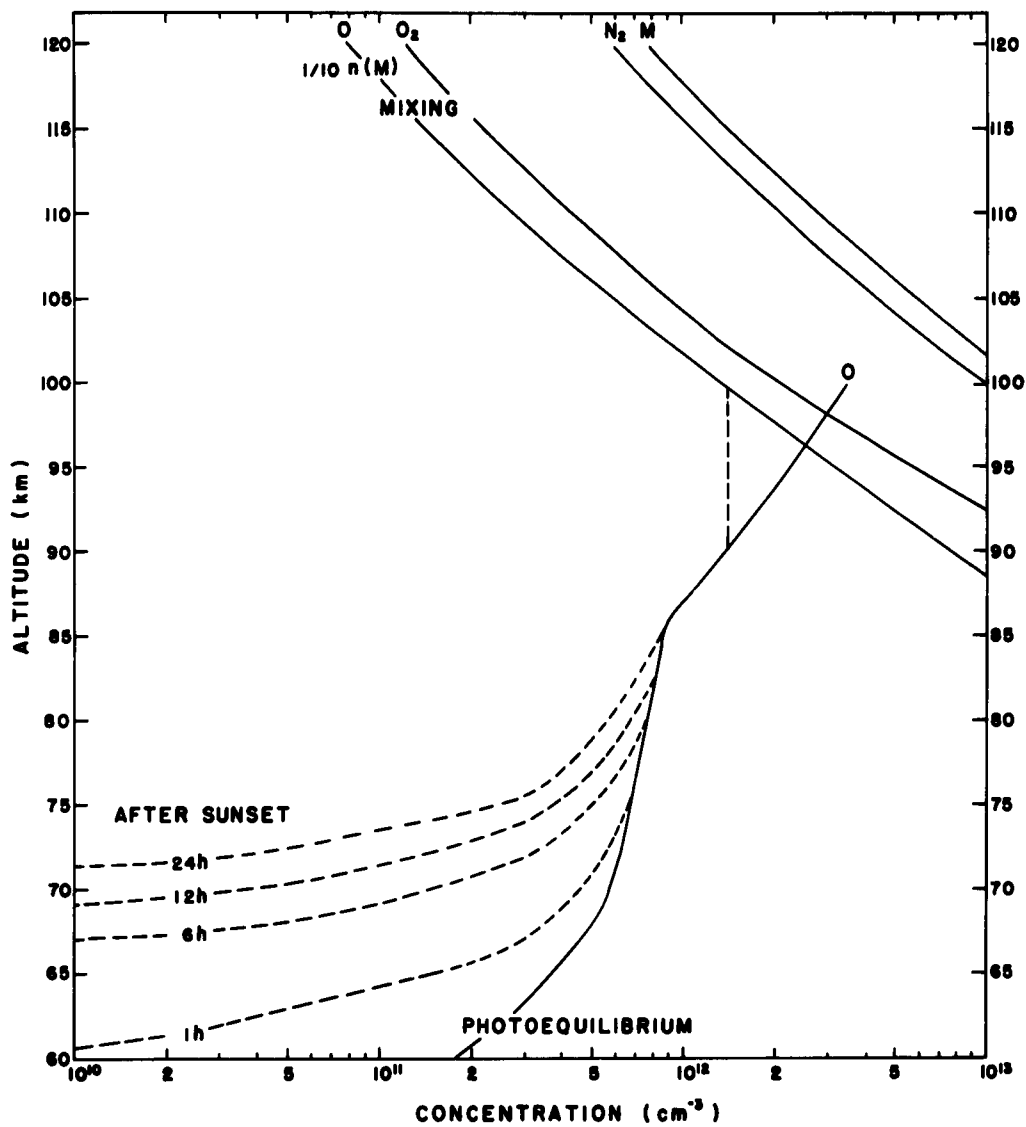
Table III
Photodissociation Rate Coefficients of O_2

Molecules (cm^{-2})	$J(O_2)$ (sec^{-1})	Molecules (cm^{-2})	$J(O_2)$ (sec^{-1})
0	4.0×10^{-6}	5×10^{18}	5.5×10^{-8}
5×10^{16}	3.0×10^{-6}	1×10^{19}	2.2×10^{-8}
1×10^{17}	2.2×10^{-6}	5×10^{19}	8.0×10^{-9}
5×10^{17}	6.0×10^{-7}	1×10^{20}	6.6×10^{-9}
1×10^{18}	3.0×10^{-7}	5×10^{20}	4.5×10^{-9}

photochemical equilibrium conditions can be obtained. The atomic oxygen concentration increases from about $2 \times 10^{11} cm^{-3}$

at 60 km to $2 \times 10^{12} \text{ cm}^{-3}$ near 95 km (see Fig. 1). After sunset, atomic oxygen disappears in the lower mesosphere by an association with molecular oxygen while in the lower thermosphere its distribution does not follow a photoequilibrium distribution. The life-times of an oxygen atom (to reduce the initial concentration to 50 percent) shown in Fig. 2 indicate that $10^{13} \text{ atoms cm}^{-3}$ are required near the mesopause level in order to reach a life time of the order of one day. In other words, since several days are involved, above the mesopause, to produce atomic oxygen concentrations of the same order as those required by photochemical equilibrium conditions, concentrations of the order of $10^{12} \text{ atoms cm}^{-3}$ (see Fig. 1 and 2) should be considered as the correct order of magnitude for the peak concentration.

There is, therefore, a transition region leading from photochemical equilibrium to mixing depending on the atmospheric conditions. The thickness of this transition region, which may begin at the mesopause, will vary according to conditions related to a vertical transport. In an atmospheric model where all conditions cannot be considered together, it is necessary to choose average conditions. Considering a transition region of the order of 10 km, i.e. from 90 km to 100 km, an average concentration of atomic oxygen corresponds to $1.4 \times 10^{12} \text{ cm}^{-3}$ and leads to $n(O) = \frac{1}{10} n(M)$ at 100 km. (see Fig. 1). Such conditions will be adopted in order to compute the vertical distribution of atomic oxygen above 100 km. But, when it is necessary to study problems in the



ATOMIC OXYGEN vs. HEIGHT

FIGURE 1

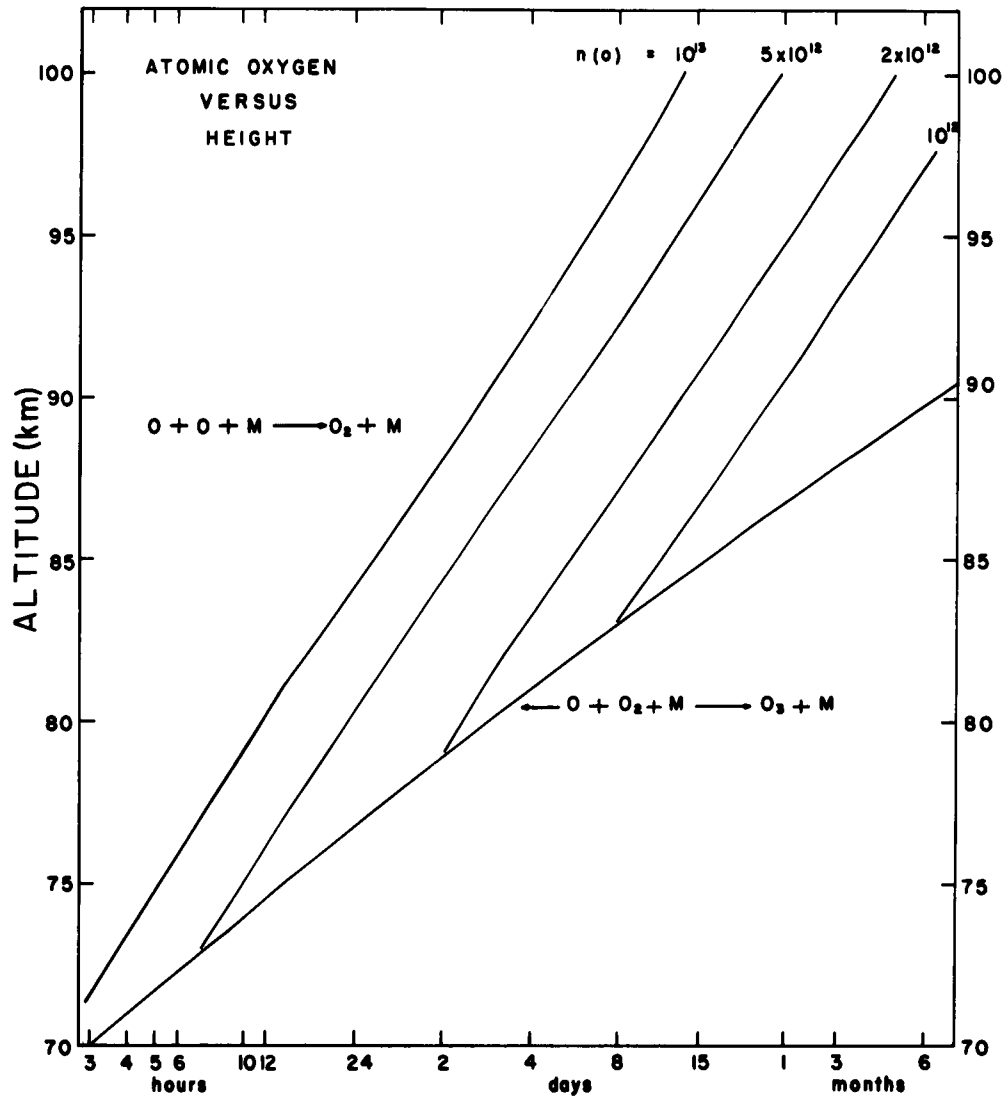


FIG. 2

LIFE TIME OF AN OXYGEN ATOM

lower thermosphere, it must be pointed out that variations will occur above the mesopause and will lead to fluctuations of the atomic oxygen concentration. Such variations must be related to atmospheric conditions which can be detected by the observations of variations in atomic or molecular processes but are not predictable. Airglow observations show how the lower thermosphere is subject to variations affecting its minor constituents.

7. Temperature Gradient in the Thermosphere

An important increase of the temperature in the thermosphere is necessary to verify the observed densities in the heterosphere even if an important increase of the atmospheric scale height can be explained by a decrease of the mean molecular mass. Two possibilities can be considered a priori: a continuous increase of the temperature related to the existence of a temperature gradient due to a permanent external heat flow and a heating by absorption of the solar radiation. The first process, suggested by Chapman (1961), was considered by Nicolet (1958) and leads to densities which are required by the observations. However, external heating cannot be the primary process since a diurnal variation of the density occurs above 200 km. The only physical possibility is the heating by ultraviolet radiation leading to strong gradients of temperature below 200 km. An important increase of the scale-height, due essentially to the increase of the temperature below 200 km, is required since it is not possible to reach the densities observed near 700 km when the temperature

gradient near 150 km is too small. In fact, the rocket observations of La Gow et al (1952, 1956, etc) reveal the presence of strong gradients of temperature below 200 km even if the absolute value of the density is not very sensitive to the temperature gradient. In other words, the boundary conditions near 100-120 km strongly affect the conditions at 200 km and the temperature gradient near 150 km determines the conditions at high altitudes (700 km).

A temperature gradient of the order of 20°K per km requires a heat transport corresponding to an energy of the order of 1 erg cm⁻² sec⁻¹ in the region where molecular nitrogen is still the principal constituent.

Since the total kinetic energy of a vertical column above 150 km is of the same order of magnitude as the ultra-violet solar energy available each day for the F region of the ionosphere, it is permissible to consider that the temperature gradient is related to the solar heating. Therefore, the general form^(*) must be as follows: the production of heat dP due to the effective solar radiation E_{uv} is written

$$dP = E_{uv} e^{-\tau} d\tau \quad (24)$$

in which τ denotes an equivalent optical depth, and the relation to the temperature is

$$T^{1/2} \frac{dT}{dz} \propto E_{uv} (1 - e^{-\tau}) \quad (25)$$

(*) see description and references in Chapter II, §2.6

"Thermal Conductivity in the Thermosphere" by M. Nicolet (1960 b)

When τ is sufficiently small, the heterosphere becomes isothermal. When $E_{uv} = 0$ (nighttime), the redistribution of temperature (isothermy) depends on the law of cooling by conduction and that the time of conduction is so short at the thermopause level that the temperature of the isothermal layer can be several hundred degrees less during the night than during the day. Thus, the thermospheric gradient varies between day and night and must be different with various conditions of solar activity.

8. Diffusion in the Thermosphere

Diffusion plays a leading role in the thermosphere. It competes with mixing and photochemical equilibrium (Nicolet and Mange, 1954; Nicolet, 1954 and 1960; Mange, 1955 and 1961). Rocket observations show that argon begins to diffuse, compared to molecular nitrogen, in the region of 110 km (Meadows and Townsend, 1958; Pokhunkov 1961). The time of diffusion τ_D can be written (Mange, 1957)

$$\tau_D = A H^2 / D_e \quad (26)$$

where A is a constant which must be determined from the continuity equation and D_e is the diffusion coefficient depending on the total concentration.

The knowledge of the exact equivalent altitude of diffusion is not very important for the total density below 200 km since molecular nitrogen remains the principal constituent. Any level between 120 km and 150 km can be used. However, when physical conditions are used for the study of the behavior of

the ionosphere and for the determination of the density above 500 km, the exact beginning of diffusion in a working model with molecular and atomic oxygen must be considered with great care.

Thus, the distribution of molecular nitrogen, which is practically unaffected by diffusion, will be almost the same in mixing and diffusion equilibrium. Since the time of diffusion of atomic and molecular oxygen does not decrease very much between 120 and 150 km, due to the rapid increase of the scale height, the level of 120 km can be taken as the beginning of diffusion to be introduced in a calculation. A change of a few kms does not change the concentrations very much. As far as the inert gases are concerned, the level may be considered below 120 km, if one considers that the absence of chemical reactions is an important factor. An altitude of (110 ± 5) km is not in disagreement with the observational evidence. As far as atomic hydrogen is concerned, the problem of diffusion is related to the effect of a source and the vertical distribution is affected by the flow which can be supported by diffusion.

9. Thermospheric Boundary Conditions

Since the first publication of Havens, Koll and LaGow (1952), there have been several measurements of the pressure near 100 km and a few measurements of the density from 120 km to 200 km. From these observational data a pressure of $(3 \pm 1) \times 10^{-4}$ mm Hg may be accepted at 100 km and a density of $(4 \pm 2) \times 10^{-13}$ gm cm⁻³ at 200 km. Since the boundary

conditions must be chosen between 100 and 120 km, examples of physical parameters are presented in Table IV. It is clear that, for the same conditions at 100 km, the various densities and concentrations at 120 km cannot be distinguished by means of observational data when the temperature changes from 262°K to 386°K, i.e a temperature difference of the order of 125°K.

However, between 150 km and 160 km, an isopycnic level can be considered for each case (a), (b) and (c). The densities at 150 km are as follows:

$$\rho(a) = (1.6 \pm 0.05) \times 10^{-12} \text{ gm cm}^{-3} \quad (27)$$

$$\rho(b) = (2.4 \pm 0.05) \times 10^{-12} \text{ gm cm}^{-3} \quad (28)$$

$$\rho(c) = (3.4 \pm 0.05) \times 10^{-12} \text{ gm cm}^{-3} \quad (29)$$

leading to values near 200 km in the range of $2 \times 10^{-13} \text{ gm cm}^{-3}$, $4 \times 10^{-13} \text{ gm cm}^{-3}$ and $6 \times 10^{-13} \text{ gm cm}^{-3}$ when the gradient of temperature is great enough at 150 km. It can be shown that it is possible to find, with the range of temperature 262°K to 386°K at 120 km, a vertical distribution of density leading to an agreement with satellite observations up to 700 km. The vertical distribution of the density between 300 km and 700 km depends on the average temperature of the whole region. However, in order to reach a density of $4 \times 10^{-13} \text{ gm cm}^{-3}$ at 200 km with a density less than $3 \times 10^{-12} \text{ gm cm}^{-3}$ at 150 km and to follow the densities obtained from the satellite observations up to 700 km, it is easy to see that the best conditions are given by the average conditions defined by case (b) of Table IV. Thus the conditions adopted at 120 km as the effective level for the calculations are as follows:

Table IV

Examples of Physical Parameters Between 100 and 120 km

Altitude (km)	100	105	110	115	120
Scale height (km)	(a) 6.37	6.87	7.37	7.87	8.37
	(b) 6.37	7.37	8.37	9.37	10.37
	(c) 6.37	7.87	9.37	10.87	12.37
Temperature (°K)	(a) 200	216	230	246	262
	(b) 200	231	262	293	324
	(c) 200	247	293	340	386
Pressure (mm Hg)	(a) 3.0×10^{-4}	1.4×10^{-4}	6.9×10^{-5}	3.7×10^{-5}	2.0×10^{-5}
	(b) 3.0×10^{-4}	1.4×10^{-4}	7.6×10^{-5}	4.3×10^{-5}	2.6×10^{-5}
	(c) 3.0×10^{-4}	1.5×10^{-4}	8.2×10^{-5}	5.0×10^{-5}	3.3×10^{-5}
Density (gm cm^{-3})	(a) 6.6×10^{-10}	2.9×10^{-10}	1.3×10^{-11}	6.5×10^{-11}	3.3×10^{-11}
	(b) 6.6×10^{-10}	2.7×10^{-10}	1.3×10^{-11}	6.5×10^{-11}	3.5×10^{-11}
	(c) 6.6×10^{-10}	2.6×10^{-10}	1.2×10^{-11}	6.5×10^{-11}	3.7×10^{-11}
Concentrations (cm^{-3})					
$n(\text{N}_2)$	(a) 1.1×10^{13}	4.7×10^{12}	2.2×10^{12}	1.0×10^{12}	5.3×10^{11}
	(b) 1.1×10^{13}	4.5×10^{12}	2.1×10^{12}	1.1×10^{11}	5.8×10^{11}
	(c) 1.1×10^{13}	4.3×10^{12}	2.0×10^{12}	1.1×10^{12}	6.1×10^{11}
$n(\text{O}_2)$	(a) 2.2×10^{12}	9.5×10^{11}	4.4×10^{11}	2.1×10^{11}	1.1×10^{11}
	(b) 2.2×10^{12}	9.2×10^{11}	4.3×10^{11}	2.2×10^{11}	1.2×10^{11}
	(c) 2.2×10^{12}	8.8×10^{11}	4.1×10^{11}	2.2×10^{11}	1.2×10^{11}

Table IV - continued

Altitude	100	105	110	115	120
Concentrations (cm^{-3})					
n(O)	(a) 1.4×10^{12}	6.1×10^{11}	2.8×10^{11}	1.4×10^{11}	7.0×10^{10}
	(b) 1.4×10^{12}	5.9×10^{11}	2.7×10^{11}	1.4×10^{11}	7.6×10^{10}
	(c) 1.4×10^{12}	5.6×10^{11}	2.6×10^{11}	1.4×10^{11}	8.0×10^{10}
n(M)	(a) 1.4×10^{13}	6.3×10^{12}	2.9×10^{12}	1.4×10^{12}	7.1×10^{11}
	(b) 1.4×10^{13}	6.0×10^{12}	2.8×10^{12}	1.4×10^{12}	7.8×10^{11}
	(c) 1.4×10^{13}	5.8×10^{12}	2.7×10^{12}	1.4×10^{12}	8.1×10^{11}

Density ρ (120 km)	= 3.5×10^{-11} gm cm ⁻³
Pressure p (120 km)	= 2.6×10^{-5} mm Hg
Temperature T (120 km)	= 324°K
Scale Height H (120 km)	= 10.37 km
Total Concentration $n(M)$	= 7.8×10^{11} cm ⁻³
Molecular Nitrogen $n(N_2)$	= 5.8×10^{11} cm ⁻³
Molecular Oxygen $n(O_2)$	= 1.2×10^{11} cm ⁻³
Atomic Oxygen $n(O)$	= 7.6×10^{10} cm ⁻³

Thus, the ratio of the atomic oxygen concentration corresponds to 25 percent dissociation of O_2 in the normal oxygen-nitrogen atmosphere.

With the preceding boundary conditions at 120 km, the isopycnic level at 150 km is well defined for a large range of scale heights. For example, if we adopt scale height gradients β between 0.3 and 0.7 from 120 km to 200 km the density at 150 km is

$$\rho_{150 \text{ km}} = (2.40 \pm 0.03) \times 10^{-12} \text{ gm cm}^{-3} \quad (30)$$

while the density at 200 km is

$$\rho_{200 \text{ km}} = (3.3 \pm 1.2) \times 10^{-13} \text{ gm cm}^{-3} \quad (31)$$

For this reason it is very convenient to start the whole computation from 150 km where the physical conditions are determined by the temperature and its gradient. At 150 km, formula (25) is applied and the equivalent optical depth is fixed in keeping the same temperature gradient in the ten kilometer ranges between 140 km and 150 km, and between 150 km and 160 km. Such a determination leads to a vertical distribution

of temperature which would correspond to a permanent heating by an overhead sun.

Several cases were considered where the conditions at 150 km are given by the gradient $\beta = dH/dz$. In order to cover all possibilities, calculations were made for 4 values of β ; namely $\beta = 0.3, 0.4, 0.5$ and 0.6 with diffusion beginning at 120 km or at 150 km. The values of the density at 200 km are as follows:

Diffusion beginning at 150 km:

$$\beta = 0.3, \quad \rho_{200}(3d) = 2.16 \times 10^{-13} \text{ gm cm}^{-3}$$

$$\beta = 0.4, \quad \rho_{200}(4d) = 2.86 \times 10^{-13} \text{ gm cm}^{-3}$$

$$\beta = 0.5, \quad \rho_{200}(5d) = 3.45 \times 10^{-13} \text{ gm cm}^{-3}$$

$$\beta = 0.6, \quad \rho_{200}(6d) = 3.92 \times 10^{-13} \text{ gm cm}^{-3}$$

Diffusion beginning at 120 km:

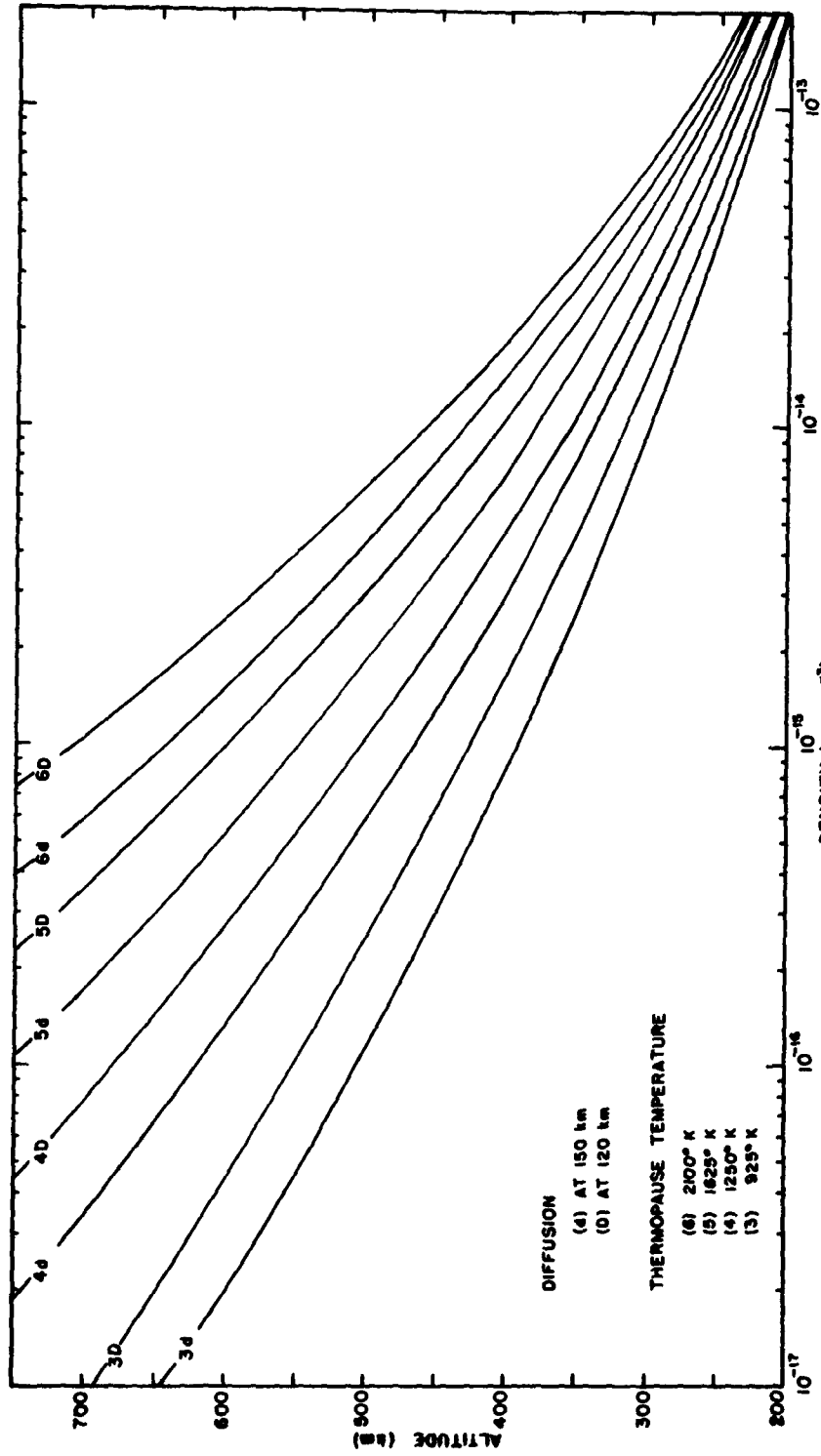
$$\beta = 0.3, \quad \rho_{200}(3D) = 2.42 \times 10^{-13} \text{ gm cm}^{-3}$$

$$\beta = 0.4, \quad \rho_{200}(4D) = 2.98 \times 10^{-13} \text{ gm cm}^{-3}$$

$$\beta = 0.5, \quad \rho_{200}(5D) = 3.62 \times 10^{-13} \text{ gm cm}^{-3}$$

$$\beta = 0.6, \quad \rho_{200}(6D) = 4.07 \times 10^{-13} \text{ gm cm}^{-3}$$

The differences in the density at 200 km are due essentially to the effect of the vertical distribution of the thermospheric temperature and not to the effect of diffusion. However, the various distributions of the density shown in Fig. 3 indicate the effect of diffusion. The d curves represent densities with diffusion beginning at 150 km which are systematically lower than the densities of D curves for diffusion beginning at 120 km. Comparison between these various curves of density and the values of density deduced from satellite



DENSITY vs HEIGHT
FIGURE 3

observations by Jacchia (1960), King-Hele and Walker (1961), Martin, Neveling, Priester and Roemer (1961), and Paetzold and Zschörner (1961) shows that the scale height gradient must reach $\beta = 0.6$ in order to explain the maximum of density which was observed in 1958. With conditions (a) and (c) of Table IV, the gradient would have been $\beta = 0.7$ and $\beta = 0.5$; respectively.

If the value of the density $\rho_{200 \text{ km}} = 4 \times 10^{-13} \text{ gm cm}^{-3}$ is accepted, we can assume, therefore, that the model 6D is a standard model able to represent the conditions of the thermosphere. For this reason, it has been chosen to represent the physical parameters of the thermosphere in which the temperature gradient varies. The thermopause must correspond to temperatures between 1900°K and 2100°K in order to reach an altitude of 500 km. Table V gives the densities, temperatures and densities at the thermopause level and also at altitudes of 400, 300 and 200 km when a cooling by conduction is applied during one hour.

The conditions at 150 km do not change. They are represented by a temperature of 877°K and a density equal to $2.4 \times 10^{-12} \text{ gm cm}^{-3}$. Table V also shows the variation of the density with temperature and height in the thermosphere when the temperature at the thermopause reaches $2000 \pm 100^{\circ}\text{K}$ and is about 500°K less at 200 km, i.e. $1500 \pm 25^{\circ}\text{K}$ at 200 km. When the temperature is less than 1900°K , the level of the thermopause decreases and a height of the order of 500 km is adopted for its maximum altitude. In this way, a considerable

Table V

Densities and Temperatures with Thermopause at 500 km

<u>500 km</u>		<u>400 km</u>		<u>300 km</u>	
T(°K)	$\rho(\text{gm.cm}^{-3})$	T	ρ	T	ρ
2100	6.6×10^{-15}	2066	1.9×10^{-14}	1968	6.7×10^{-14}
2070	6.4×10^{-15}	2045	1.9×10^{-14}	1945	6.7×10^{-14}
2045	6.2×10^{-15}	2023	1.8×10^{-14}	1924	6.7×10^{-14}
2020	6.0×10^{-15}	2000	1.8×10^{-14}	1904	6.6×10^{-14}
1995	5.9×10^{-15}	1978	1.7×10^{-14}	1884	6.6×10^{-14}
1975	5.7×10^{-15}	1956	1.7×10^{-14}	1864	6.5×10^{-14}
1900	5.2×10^{-15}	1890	1.6×10^{-14}	1807	6.3×10^{-14}
1837*	4.8×10^{-15}	1826	1.5×10^{-14}	1752	6.2×10^{-14}

(*) Thermopause at 460 km.

200 km

T	ρ
1530	4.1×10^{-13}
1520	4.1×10^{-13}
1510	4.1×10^{-13}
1497	4.1×10^{-13}
1486	4.1×10^{-13}
1475	4.1×10^{-13}
1444	4.1×10^{-13}
1414	4.1×10^{-13}

part of the heterosphere can be studied as an isothermal atmosphere. A vertical isothermy is justified since the increase of temperature between 300 km and 500 km is unimportant compared with its increase in the thermosphere below 300 km. Thus, an atmospheric model in which the level of the thermopause decreases when its temperature decreases, leads to standard conditions which can represent the effect of solar activity as well as a diurnal variation.

10. Atmospheric Parameters from 120 km to 500 km

Since the absolute value of density in the upper part of the thermosphere depends on the absolute value of the temperature, and is related to its gradient near 150 km, the calculation was made by introducing various gradients consistent with a cooling by conduction of the whole atmosphere and considered as a plane-parallel atmosphere in diffusive equilibrium. Starting from the model 6D represented by Table V, we can adopt the following conditions between 150 km and 160 km:

No.	6D	1	1.5	2	3	4	5	6	7
T(150 km)	877	873	864	852	829	803	777	751	726
(dT/dz)km	18	15	14	13	11	10	9	8	7

Such conditions lead to a large range of variation of the physical parameters corresponding to a difference of more than 1000°K in the temperature of the thermopause. The vertical distributions of the temperature given in Table VI show

how they are related to the altitude of the thermopause. In other words, a diurnal effect corresponds to a decrease of the temperature associated with a decrease of the altitude of the thermopause. Any increase of the heating due to solar activity must be related to an increase of the altitude of the thermopause and of its temperature. Other physical parameters of the atmosphere between 120 km and 500 km corresponding to the data of Table VI are given in the following Tables (VII to XV).

Table VII - Density in gm cm^{-3} due to molecular nitrogen, and atomic and molecular oxygen in diffusive equilibrium between 120 km and 500 km. The variation of the density at 500 km corresponds to a factor of the order of 10. It can be seen that there is a linear relationship between the temperature and the altitude of surfaces of equal density.

Table VIII - Mean molecular mass in units of mass of atomic oxygen; $M = 16$. The mass $M = 27.4$ at 120 km shows the extent of dissociation of oxygen in a molecular oxygen - nitrogen atmosphere; $M = 29$. The decrease from $M = 27.4$ to about $M = 16$ corresponds to the increasing importance with height of atomic oxygen compared with molecular nitrogen; $M = 28$. It is clear that the mean molecular mass does not depend strongly on the temperature in the thermosphere and must be determined with very great precision if it is used as an observational parameter.

Table IX - Total concentration in cm^{-3} . If the total concentration remains practically constant at 200 km, its variation at 500 km attains a factor of 10. It should be pointed out

TABLE VI - Temperature (°K)

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	324	324	324	324	324	324	324	324	324
130	509	509	509	509	509	509	509	509	509
140	694	694	696	695	688	677	664	649	634
150	877	873	863	852	829	803	777	751	726
160	1059	1022	1002	982	941	902	864	827	793
170	1212	1148	1117	1088	1031	979	930	884	842
180	1340	1252	1212	1174	1104	1039	981	927	878
190	1448	1340	1292	1246	1162	1087	1020	960	905
200	1540	1414	1358	1305	1210	1126	1051	985	925
210	1618	1477	1413	1355	1248	1156	1075	1004	940
220	1685	1530	1460	1396	1280	1180	1094	1018	951
240	1791	1613	1532	1458	1327	1215	1120	1037	965
260	1870	1674	1585	1503	1358	1237	1136	1048	973
280	1930	1719	1622	1534	1379	1252	1145	1054	977
300	1975	1752	1650	1556	1393	1261	1150	1057	978
320	2010	1777	1670	1572	1402	1267	1153	1059	
340	2036	1796	1685	1583	1408	1271	1155		
360	2057	1809	1695	1590	1411	1272			
380	2073	1819	1702	1594	1412				
400	2086	1826	1707	1597					
420	2096	1831	1710	1598					
440	2104	1835	1711						
460	2110	1837							
480	2115								
500	2119								

TABLE VII - Density (gm cm^{-3}) Oxygen and Nitrogen

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}	3.54×10^{-11}
130	1.04	1.04	1.04	1.04	1.04	1.04	1.04	1.04	1.02
140	4.52×10^{-12}	4.52×10^{-12}	4.52×10^{-12}	4.52×10^{-12}	4.59×10^{-12}	4.65×10^{-12}	4.71×10^{-12}	4.76×10^{-12}	4.71×10^{-12}
150	2.41	2.43	2.45	2.48	2.54	2.59	2.62	2.65	2.61
160	1.46	1.50	1.52	1.54	1.58	1.61	1.63	1.64	1.60
170	9.79×10^{-13}	1.01	1.03	1.04	1.07	1.08	1.08	1.08	1.05
180	7.01	7.26×10^{-13}	7.35×10^{-13}	7.42×10^{-13}	7.54×10^{-13}	7.58×10^{-13}	7.54×10^{-13}	7.44×10^{-13}	7.14×10^{-13}
190	5.25	5.40	5.45	5.48	5.53	5.51	5.42	5.28	5.00
200	4.07	4.14	4.17	4.17	4.16	4.10	3.98	3.84	3.59
210	3.22	3.25	3.25	3.24	3.20	3.11	2.98	2.83	2.61
220	2.61	2.60	2.58	2.56	2.50	2.39	2.27	2.12	1.93
240	1.77	1.72	1.70	1.66	1.58	1.47	1.35	1.23	1.09
260	1.26	1.19	1.16	1.11	1.03	9.35×10^{-14}	8.36×10^{-14}	7.40×10^{-14}	6.35×10^{-14}
280	9.20×10^{-14}	8.47×10^{-14}	8.10×10^{-14}	7.71×10^{-14}	6.93×10^{-14}	6.11	5.31	4.57	3.81
300	6.86	6.15	5.80	5.44	4.76	4.08	3.45	2.89	2.35
320	5.20	4.54	4.23	3.91	3.32	2.77	2.29	1.87	1.48
340	4.00	3.40	3.13	2.86	2.36	1.92	1.54	1.23	9.55×10^{-15}
360	3.11	2.58	2.34	2.11	1.70	1.34	1.06	8.23×10^{-15}	6.26
380	2.44	1.98	1.77	1.58	1.24	9.56×10^{-15}	7.34×10^{-15}	5.60	4.16
400	1.93	1.53	1.36	1.19	9.11×10^{-15}	6.89	5.16	3.86	2.81
420	1.55	1.20	1.04	9.07×10^{-15}	6.78	5.01	3.68	2.69	1.92
440	1.24	9.42×10^{-15}	8.13×10^{-15}	6.97	5.08	3.68	2.64	1.89	1.33
460	1.00	7.46	6.37	5.40	3.84	2.73	1.92	1.35	9.24×10^{-16}
480	8.14×10^{-15}	5.95	5.02	4.21	2.93	2.04	1.40	9.64×10^{-16}	6.48
500	6.65	4.78	3.98	3.30	2.25	1.53	1.03	6.95	4.58

TABLE VIII - Mean Molecular Mass (Atomic Oxygen = 16), Oxygen and Nitrogen

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	27.4	27.4	27.4	27.4	27.4	27.4	27.4	27.4	27.4
130	26.9	26.9	26.9	26.9	26.9	26.9	26.9	26.9	26.9
140	26.5	26.5	26.5	26.5	26.5	26.5	26.5	26.5	26.5
150	26.2	26.2	26.2	26.2	26.2	26.1	26.1	26.1	26.1
160	25.8	25.8	25.8	25.8	25.8	25.8	25.8	25.7	25.7
170	25.6	25.5	25.5	25.5	25.5	25.4	25.4	25.3	25.3
180	25.3	25.3	25.2	25.2	25.2	25.1	25.0	24.9	24.8
190	25.1	25.0	25.0	24.9	24.8	24.8	24.7	24.5	24.4
200	24.8	24.7	24.7	24.6	24.5	24.4	24.3	24.1	24.0
210	24.6	24.5	24.4	24.3	24.2	24.1	23.9	23.7	23.5
220	24.3	24.2	24.1	24.0	23.9	23.7	23.5	23.3	23.1
240	23.9	23.7	23.6	23.5	23.3	23.0	22.8	22.5	22.2
260	23.4	23.2	23.0	22.9	22.6	22.3	22.0	21.7	21.3
280	23.0	22.7	22.5	22.4	22.0	21.6	21.3	20.9	20.5
300	22.6	22.2	22.0	21.8	21.4	21.0	20.6	20.2	19.8
320	22.2	21.8	21.5	21.3	20.8	20.4	19.9	19.5	19.1
340	21.8	21.3	21.0	20.8	20.3	19.8	19.4	18.9	18.5
360	21.4	20.8	20.6	20.3	19.8	19.3	18.8	18.4	18.0
380	21.0	20.4	20.1	19.9	19.3	18.8	18.4	18.0	17.6
400	20.6	20.0	19.7	19.4	18.9	18.4	18.0	17.6	17.3
420	20.3	19.7	19.4	19.1	18.5	18.0	17.6	17.3	17.2
440	19.9	19.3	19.0	18.7	18.2	17.7	17.3	17.0	16.8
460	19.6	19.0	18.7	18.4	17.9	17.5	17.1	16.8	16.6
480	19.3	18.7	18.4	18.1	17.6	17.2	16.9	16.7	16.5
500	19.0	18.4	18.1	17.9	17.4	16.9	16.6	16.5	16.4

TABLE IX - Total Concentration (cm^{-3}) Oxygen and nitrogen

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	7.8×10^{11}	7.8×10^{11}	7.8×10^{11}	7.8×10^{11}	7.8×10^{11}	7.8×10^{11}	7.8×10^{11}	7.8×10^{11}	7.6×10^{11}
130	2.3	2.3	2.3	2.3	2.3	2.3	2.3	2.3	2.3
140	1.0	1.0	1.0	1.0	1.0	1.0	1.1	1.1	1.1
150	5.6×10^{10}	5.6×10^{10}	5.6×10^{10}	5.7×10^{10}	5.8×10^{10}	6.0×10^{10}	6.0×10^{10}	6.1×10^{10}	6.0×10^{10}
160	3.4	3.5	3.6	3.6	3.7	3.8	3.8	3.8	3.8
170	2.3	2.4	2.4	2.4	2.5	2.6	2.6	2.6	2.5
180	1.7	1.7	1.8	1.8	1.8	1.8	1.8	1.8	1.7
190	1.3	1.3	1.3	1.3	1.3	1.3	1.3	1.3	1.2
200	9.9×10^9	1.0	1.0	1.0	1.0	1.0	9.9×10^9	9.6×10^9	9.0×10^9
210	7.9	8.0×10^9	8.0×10^9	8.0×10^9	8.0×10^9	7.8×10^9	7.5	7.2	6.7
220	6.4	6.5	6.4	6.4	6.3	6.1	5.8	5.5	5.0
240	4.5	4.4	4.3	4.2	4.1	3.8	3.6	3.3	2.9
260	3.2	3.1	3.0	2.9	2.7	2.5	2.3	2.0	1.8
280	2.4	2.2	2.2	2.1	1.9	1.7	1.5	1.3	1.1
300	1.8	1.7	1.6	1.5	1.3	1.2	1.0	8.6×10^8	7.2×10^8
320	1.4	1.2	1.2	1.1	9.6×10^8	8.2×10^8	6.9×10^8	5.8	4.7
340	1.1	9.6×10^8	8.9×10^8	8.3×10^8	7.0	5.8	4.8	3.9	3.1
360	8.8×10^8	7.5	6.8	6.2	5.2	4.2	3.4	2.7	2.1
380	7.0	5.8	5.3	4.8	3.8	3.1	2.4	1.9	1.4
400	5.6	4.6	4.1	3.7	2.9	2.2	1.7	1.3	9.8×10^7
420	4.6	3.7	3.2	2.9	2.2	1.7	1.2	9.4×10^7	6.8
440	3.7	2.9	2.6	2.2	1.7	1.2	9.2×10^7	6.7	4.8
460	3.1	2.4	2.0	1.8	1.3	9.4×10^7	6.8	4.8	3.3
480	2.5	1.9	1.6	1.4	1.0	7.1	5.0	3.5	2.4
500	2.1	1.6	1.3	1.1	7.8×10^7	5.4	3.7	2.5	1.7

that a factor of the order 100 represents the decrease of the first hundred km above 120 km and of about 300 km above 200 km for a median model of this Table.

Table X - Pressure in mm Hg. The variation of the pressure represents the effect of the total variation of the concentrations and temperatures given in Tables VI and IX corresponding to a variation of a factor of the order of 25 at 500 km.

Table XI - Scale heights in km. The variation of the ratio of temperature and mean molecular mass is given by values of the scale heights ranging between 10 and 100 km for all atmospheric conditions between 120 km and 500 km.

Table XII - Atomic oxygen concentrations in cm^{-3} . The concentration at 120 km corresponds to a percentage of oxygen dissociation equal to 25%. The diffusion distribution leads to concentrations between 2×10^7 and 2×10^8 atoms cm^{-3} at 500 km for temperatures between 1000°K and 2000°K ; respectively.

Table XIII - Molecular nitrogen concentration in cm^{-3} . The effect of the diffusion is such that the concentration of molecular nitrogen at 500 km cannot reach 10 percent of that of atomic oxygen if the temperature is less than 1600°K . The molecular nitrogen concentration is very sensitive to the temperature since its variation at 500 km corresponds to about a factor of 100.

Table XIV - Molecular oxygen concentrations in cm^{-3} . A value of about $6 \times 10^9 \text{ cm}^{-3}$ at 150 km is justified by observational

TABLE X - Pressure (mm Hg)

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	2.6×10^{-5}	2.6×10^{-5}	2.6×10^{-5}	2.6×10^{-5}	2.6×10^{-5}	2.6×10^{-5}	2.6×10^{-5}	2.6×10^{-5}	2.5×10^{-5}
130	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.2
140	7.4×10^{-6}	7.4×10^{-6}	7.4×10^{-6}	7.4×10^{-6}	7.4×10^{-6}	7.4×10^{-6}	7.4×10^{-6}	7.3×10^{-6}	7.0×10^{-6}
150	5.0	5.0	5.0	5.0	5.0	5.0	4.9	4.8	4.5
160	3.7	3.7	3.7	3.6	3.6	3.5	3.4	3.3	3.1
170	2.9	2.8	2.8	2.8	2.7	2.6	2.5	2.4	2.2
180	2.3	2.2	2.2	2.2	2.1	2.0	1.8	1.7	1.6
190	1.9	1.8	1.8	1.7	1.6	1.5	1.4	1.3	1.2
200	1.6	1.5	1.4	1.4	1.3	1.2	1.1	9.8×10^{-7}	8.6×10^{-7}
210	1.3	1.2	1.2	1.1	1.0	9.3×10^{-7}	8.4×10^{-7}	7.5	6.5
220	1.1	1.0	9.7×10^{-7}	9.2×10^{-7}	8.3×10^{-7}	7.4	6.6	5.8	5.0
240	8.3×10^{-7}	7.3×10^{-7}	6.9	6.4	5.6	4.8	4.2	3.5	2.9
260	6.2	5.4	5.0	4.6	3.8	3.2	2.7	2.2	1.8
280	4.8	4.0	3.6	3.3	2.7	2.2	1.8	1.4	1.1
300	3.7	3.0	2.7	2.4	1.9	1.5	1.2	9.4×10^{-8}	7.2×10^{-8}
320	2.9	2.3	2.0	1.8	1.4	1.1	8.2×10^{-8}	6.3	4.7
340	2.3	1.8	1.6	1.4	1.0	7.7×10^{-8}	5.7	4.3	3.1
360	1.9	1.4	1.2	1.0	7.5×10^{-8}	5.5	4.0	3.0	2.1
380	1.5	1.1	9.3×10^{-8}	7.9×10^{-8}	5.6	4.0	2.9	2.0	1.4
400	1.2	8.7×10^{-8}	7.3	6.1	4.2	3.0	2.1	1.4	9.9×10^{-9}
420	1.0	7.0	5.8	4.7	3.2	2.2	1.5	1.0	6.9
440	8.2×10^{-8}	5.6	4.6	3.7	2.5	1.6	1.1	7.3×10^{-9}	4.8
460	6.7	4.5	3.6	2.9	1.9	1.2	8.1×10^{-9}	5.3	3.4
480	5.6	3.6	2.9	2.3	1.5	9.4×10^{-9}	6.0	3.8	2.4
500	4.7	3.0	2.4	1.9	1.2	7.4	4.7	3.0	1.9

TABLE XI - Atmospheric scale height (km) Oxygen and nitrogen

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	10.4	10.4	10.4	10.4	10.4	10.4	10.4	10.4	10.4
130	16.7	16.7	16.7	16.7	16.7	16.7	16.7	16.7	16.5
140	23.2	23.2	23.2	23.2	23.0	22.6	22.2	21.7	21.2
150	29.8	29.6	29.3	28.9	28.1	27.3	26.4	25.6	24.7
160	36.5	35.2	34.6	33.8	32.5	31.1	29.9	28.7	27.5
170	42.4	40.1	39.1	38.1	36.2	34.4	32.7	31.2	29.8
180	47.4	44.4	43.0	41.7	39.3	37.1	35.1	33.3	31.7
190	52.0	48.2	46.5	44.9	42.0	39.5	37.2	35.2	33.3
200	56.0	51.6	49.6	47.8	44.5	41.6	39.0	36.8	34.8
210	59.6	54.6	52.4	50.3	46.6	43.4	40.7	38.3	36.1
220	62.8	57.3	54.9	52.6	48.6	45.2	42.2	39.6	37.4
240	68.4	62.1	59.3	56.7	52.0	48.2	44.9	42.1	39.7
260	73.3	66.3	63.1	60.2	55.1	50.9	47.4	44.4	41.9
280	77.5	70.0	66.5	63.4	57.9	53.4	49.7	46.6	44.0
300	81.3	73.3	69.6	66.3	60.5	55.8	52.0	48.7	46.0
320	84.8	76.4	72.6	69.0	62.9	58.1	54.1	50.7	47.9
340	88.0	79.4	75.3	71.6	65.2	60.3	56.1	52.6	49.7
360	91.1	82.1	77.9	74.0	67.5	62.4	58.0	54.4	51.3
380	94.1	84.8	80.4	76.4	69.6	64.4	59.8	56.1	52.8
400	96.9	87.3	82.8	78.6	71.6	66.2	61.5	57.6	54.2
420	99.6	89.7	85.1	80.8	73.5	68.0	63.1	59.0	55.4
440	102.3	92.0	87.2	82.8	75.3	69.6	64.5	60.2	56.4
460	104.9	94.3	89.3	84.7	77.0	71.1	65.8	61.3	57.3
480	107.4	96.4	91.2	86.6	78.7	72.5	67.0	62.3	58.2
500	110.8	98.4	93.1	88.3	80.2	73.8	68.0	63.1	58.9

TABLE XII - Atomic Oxygen Concentration (cm^{-3})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	7.6×10^{10}	7.6×10^{10}	7.6×10^{10}	7.6×10^{10}	7.6×10^{10}	7.6×10^{10}	7.6×10^{10}	7.6×10^{10}	7.4×10^{11}
130	3.1	3.1	3.1	3.1	3.1	3.1	3.1	3.1	3.0
140	1.7	1.7	1.7	1.7	1.7	1.7	1.8	1.8	1.8
150	1.0	1.1	1.1	1.1	1.1	1.1	1.2	1.2	1.2
160	7.2×10^9	7.5×10^9	7.6×10^9	7.7×10^9	8.0×10^9	8.2×10^9	8.4×10^9	8.6×10^9	8.5×10^9
170	5.4	5.6	5.8	5.9	6.1	6.2	6.4	6.5	6.5
180	4.2	4.5	4.6	4.6	4.8	4.9	5.0	5.1	5.0
190	3.5	3.6	3.7	3.8	3.9	4.0	4.0	4.1	4.0
200	2.9	3.0	3.1	3.1	3.2	3.3	3.3	3.3	3.2
210	2.5	2.6	2.6	2.6	2.7	2.7	2.7	2.7	2.6
220	2.1	2.2	2.2	2.2	2.3	2.3	2.3	2.2	2.1
240	1.6	1.7	1.7	1.7	1.7	1.7	1.6	1.6	1.5
260	1.3	1.3	1.3	1.3	1.3	1.2	1.2	1.1	1.0
280	1.0	1.0	1.0	1.0	9.7×10^8	9.2×10^8	8.6×10^8	7.9×10^8	7.1×10^8
300	8.6×10^8	8.3×10^8	8.1×10^8	7.9×10^8	7.5	6.9	6.3	5.7	5.0
320	7.1	6.7	6.5	6.3	5.8	5.3	4.7	4.1	3.5
340	5.9	5.5	5.3	5.0	4.6	4.0	3.5	3.0	2.5
360	5.0	4.5	4.3	4.0	3.6	3.1	2.6	2.2	1.7
380	4.2	3.7	3.5	3.3	2.8	2.4	1.9	1.6	1.2
400	3.5	3.1	2.9	2.6	2.2	1.8	1.4	1.1	8.8×10^7
420	3.0	2.6	2.4	2.1	1.8	1.4	1.1	8.4×10^7	6.2
440	2.5	2.1	1.9	1.7	1.4	1.1	8.2×10^7	6.1	4.4
460	2.2	1.8	1.6	1.4	1.1	8.3×10^7	6.1	4.5	3.2
480	1.8	1.5	1.3	1.2	8.7×10^7	6.4	4.6	3.3	2.3
500	1.6	1.2	1.1	9.4×10^7	6.9	5.0	3.5	2.4	1.6

TABLE XIII - Molecular Nitrogen Concentration (cm^{-3})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.7×10^{11}
130	1.7	1.7	1.7	1.7	1.7	1.7	1.7	1.7	1.7
140	7.4×10^{10}	7.4×10^{10}	7.3×10^{10}	7.4×10^{10}	7.5×10^{10}	7.6×10^{10}	7.6×10^{10}	7.7×10^{10}	7.6×10^{10}
150	3.9	3.9	3.9	4.0	4.1	4.2	4.2	4.2	4.2
160	2.3	2.4	2.4	2.4	2.5	2.6	2.6	2.6	2.5
170	1.5	1.6	1.6	1.6	1.7	1.7	1.7	1.7	1.6
180	1.1	1.1	1.1	1.1	1.2	1.2	1.2	1.1	1.1
190	8.1×10^9	8.3×10^9	8.3×10^9	8.4×10^9	8.4×10^9	8.3×10^9	8.1×10^9	7.9×10^9	7.4×10^9
200	6.2	6.2	6.3	6.2	6.2	6.1	5.8	5.6	5.2
210	4.8	4.8	4.8	4.8	4.7	4.5	4.3	4.0	3.6
220	3.8	3.8	3.8	3.7	3.6	3.4	3.2	2.9	2.6
240	2.5	2.4	2.4	2.3	2.1	2.0	1.8	1.6	1.3
260	1.7	1.6	1.6	1.5	1.3	1.2	1.0	8.6×10^8	7.1×10^8
280	1.2	1.1	1.0	9.7×10^8	8.4×10^8	7.1×10^8	5.9×10^8	4.8	3.8
300	8.8×10^8	7.6×10^8	7.0×10^8	6.5	5.4	4.4	3.5	2.7	2.0
320	6.4	5.3	4.8	4.4	3.5	2.7	2.0	1.5	1.1
340	4.7	3.8	3.4	3.0	2.3	1.7	1.2	8.7×10^7	6.0×10^7
360	3.5	2.7	2.4	2.0	1.5	1.0	7.3×10^7	5.0	3.3
380	2.6	1.9	1.6	1.4	9.8×10^7	6.6×10^7	4.4	2.8	1.8
400	1.9	1.4	1.2	9.7×10^7	6.4	4.2	2.6	1.6	9.8×10^6
420	1.5	1.0	8.3×10^7	6.7	4.3	2.6	1.6	9.4×10^6	5.4
440	1.1	7.4×10^7	5.9	4.7	2.8	1.7	9.6×10^6	5.5	3.0
460	8.4×10^7	5.4	4.2	3.2	1.9	1.1	5.9	3.2	1.7
480	6.4	4.0	3.0	2.3	1.2	6.8×10^6	3.6	1.8	9.2×10^5
500	4.8	2.9	2.2	1.6	8.4×10^6	4.4	2.2	1.1	5.2

TABLE XIV - Molecular Oxygen Concentration (maximum value) (cm^{-3})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}
130	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.0×10^{10}
140	1.2	1.2	1.2	1.2	1.3	1.3	1.3	1.3	1.3
150	6.2×10^9	6.2×10^9	6.3×10^9	6.3×10^9	6.5×10^9	6.6×10^9	6.7×10^9	6.7×10^9	6.6×10^9
160	3.5	3.6	3.7	3.7	3.8	3.8	3.9	3.9	3.8
170	2.2	2.3	2.4	2.4	2.4	2.4	2.4	2.4	2.3
180	1.5	1.6	1.6	1.6	1.6	1.6	1.6	1.5	1.4
190	1.1	1.1	1.1	1.1	1.1	1.1	1.0	1.0	9.4×10^8
200	8.2×10^8	8.2×10^8	8.2×10^8	8.1×10^8	8.0×10^8	7.7×10^8	7.3×10^8	6.8×10^8	6.2
210	6.2	6.2	6.1	6.0	5.8	5.5	5.1	4.7	4.2
220	4.8	4.7	4.6	4.5	4.3	4.0	3.6	3.3	2.8
240	3.0	2.8	2.7	2.6	2.4	2.1	1.9	1.6	1.3
260	2.0	1.8	1.7	1.6	1.4	1.2	1.0	8.2×10^7	6.5×10^7
280	1.3	1.2	1.1	9.9×10^7	8.2×10^7	6.7×10^7	5.4×10^7	4.2	3.2
300	9.2×10^7	7.6×10^7	6.9×10^7	6.2	5.0	3.8	2.9	2.2	1.6
320	6.4	5.1	4.5	4.0	3.0	2.2	1.6	1.1	7.8×10^6
340	4.5	3.4	3.0	2.6	1.8	1.3	8.9×10^6	6.0×10^6	3.9
360	3.2	2.3	2.0	1.7	1.1	7.6×10^6	4.9	3.1	1.9
380	2.3	1.6	1.3	1.1	7.1×10^6	4.5	2.8	1.7	9.7×10^5
400	1.6	1.1	8.9×10^6	7.1×10^6	4.4	2.6	1.5	8.8×10^5	4.9
420	1.2	7.7×10^6	6.0	4.7	2.7	1.6	8.7×10^5	4.7	2.5
440	8.6×10^6	5.3	4.1	3.1	1.7	9.3×10^5	4.9	2.5	1.2
460	6.3	3.7	2.8	2.0	1.1	5.5	2.8	1.4	6.4×10^4
480	4.6	2.6	1.9	1.4	6.8×10^5	3.3	1.6	7.3×10^4	3.3
500	3.4	1.8	1.3	9.0×10^5	4.3	2.0	8.9×10^4	3.9	1.7

TABLE XIII - Molecular Nitrogen Concentration (cm^{-3})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.8×10^{11}	5.7×10^{11}
130	1.7	1.7	1.7	1.7	1.7	1.7	1.7	1.7	1.7
140	7.4×10^{10}	7.4×10^{10}	7.3×10^{10}	7.4×10^{10}	7.5×10^{10}	7.6×10^{10}	7.6×10^{10}	7.7×10^{10}	7.6×10^{10}
150	3.9	3.9	3.9	4.0	4.1	4.2	4.2	4.2	4.2
160	2.3	2.4	2.4	2.4	2.5	2.6	2.6	2.6	2.5
170	1.5	1.6	1.6	1.6	1.7	1.7	1.7	1.7	1.6
180	1.1	1.1	1.1	1.1	1.2	1.2	1.2	1.1	1.1
190	8.1×10^9	8.3×10^9	8.3×10^9	8.4×10^9	8.4×10^9	8.3×10^9	8.1×10^9	7.9×10^9	7.4×10^9
200	6.2	6.2	6.3	6.2	6.2	6.1	5.8	5.6	5.2
210	4.8	4.8	4.8	4.8	4.7	4.5	4.3	4.0	3.6
220	3.8	3.8	3.8	3.7	3.6	3.4	3.2	2.9	2.6
240	2.5	2.4	2.4	2.3	2.1	2.0	1.8	1.6	1.3
260	1.7	1.6	1.6	1.5	1.3	1.2	1.0	8.6×10^8	7.1×10^8
280	1.2	1.1	1.0	9.7×10^8	8.4×10^8	7.1×10^8	5.9×10^8	4.8	3.8
300	8.8×10^8	7.6×10^8	7.0×10^8	6.5	5.4	4.4	3.5	2.7	2.0
320	6.4	5.3	4.8	4.4	3.5	2.7	2.0	1.5	1.1
340	4.7	3.8	3.4	3.0	2.3	1.7	1.2	8.7×10^7	6.0×10^7
360	3.5	2.7	2.4	2.0	1.5	1.0	7.3×10^7	5.0	3.3
380	2.6	1.9	1.6	1.4	9.8×10^7	6.6×10^7	4.4	2.8	1.8
400	1.9	1.4	1.2	9.7×10^7	6.4	4.2	2.6	1.6	9.8×10^6
420	1.5	1.0	8.3×10^7	6.7	4.3	2.6	1.6	9.4×10^6	5.4
440	1.1	7.4×10^7	5.9	4.7	2.8	1.7	9.6×10^6	5.5	3.0
460	8.4×10^7	5.4	4.2	3.2	1.9	1.1	5.9	3.2	1.7
480	6.4	4.0	3.0	2.3	1.2	6.8×10^6	3.6	1.8	9.2×10^5
500	4.8	2.9	2.2	1.6	8.4×10^6	4.4	2.2	1.1	5.2

TABLE XIV - Molecular Oxygen Concentration (maximum value) (cm^{-3})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}	1.2×10^{11}
130	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.1×10^{10}	3.0×10^{10}
140	1.2	1.2	1.2	1.2	1.3	1.3	1.3	1.3	1.3
150	6.2×10^9	6.2×10^9	6.3×10^9	6.3×10^9	6.5×10^9	6.6×10^9	6.7×10^9	6.7×10^9	6.6×10^9
160	3.5	3.6	3.7	3.7	3.8	3.8	3.9	3.9	3.8
170	2.2	2.3	2.4	2.4	2.4	2.4	2.4	2.4	2.3
180	1.5	1.6	1.6	1.6	1.6	1.6	1.6	1.5	1.4
190	1.1	1.1	1.1	1.1	1.1	1.1	1.0	1.0	9.4×10^8
200	8.2×10^8	8.2×10^8	8.2×10^8	8.1×10^8	8.0×10^8	7.7×10^8	7.3×10^8	6.8×10^8	6.2
210	6.2	6.2	6.1	6.0	5.8	5.5	5.1	4.7	4.2
220	4.8	4.7	4.6	4.5	4.3	4.0	3.6	3.3	2.8
240	3.0	2.8	2.7	2.6	2.4	2.1	1.9	1.6	1.3
260	2.0	1.8	1.7	1.6	1.4	1.2	1.0	8.2×10^7	6.5×10^7
280	1.3	1.2	1.1	9.9×10^7	8.2×10^7	6.7×10^7	5.4×10^7	4.2	3.2
300	9.2×10^7	7.6×10^7	6.9×10^7	6.2	5.0	3.8	2.9	2.2	1.6
320	6.4	5.1	4.5	4.0	3.0	2.2	1.6	1.1	7.8×10^6
340	4.5	3.4	3.0	2.6	1.8	1.3	8.9×10^6	6.0×10^6	3.9
360	3.2	2.3	2.0	1.7	1.1	7.6×10^6	4.9	3.1	1.9
380	2.3	1.6	1.3	1.1	7.1×10^6	4.5	2.8	1.7	9.7×10^5
400	1.6	1.1	8.9×10^6	7.1×10^6	4.4	2.6	1.5	8.8×10^5	4.9
420	1.2	7.7×10^6	6.0	4.7	2.7	1.6	8.7×10^5	4.7	2.5
440	8.6×10^6	5.3	4.1	3.1	1.7	9.3×10^5	4.9	2.5	1.2
460	6.3	3.7	2.8	2.0	1.1	5.5	2.8	1.4	6.4×10^4
480	4.6	2.6	1.9	1.4	6.8×10^5	3.3	1.6	7.3×10^4	3.3
500	3.4	1.8	1.3	9.0×10^5	4.3	2.0	8.9×10^4	3.9	1.7

data. However, the vertical distribution, which is assumed to be in complete diffusion equilibrium, can be affected when important diurnal variations are involved. This explains why 'maximum values' are indicated.

Table XV - $\rho H^{1/2}$ in $\text{gm cm}^{-5/2}$. This parameter is given since it is adopted in a first approximation as being proportional to the rate of change of period of a satellite.

The density ρ is taken from Table VII and H from Table XI.

Table XVI - Scale height of atomic oxygen in km. The value of the scale height corresponds to the temperatures of Table VI and mass $M = 16$. The corresponding values of molecular oxygen are given by the values of this Table divided by 2.

Table XVII - Scale height of molecular nitrogen in km. The mass is $M = 28$.

Table XVIII - $n(O) H(O) \text{ cm}^{-2}$.

Table XIX - $n(N_2) H(N_2) \text{ cm}^{-2}$.

Table XX - $n(O_2) H(O_2) \text{ cm}^{-2}$.

11. Heterosphere Above 500 km

The mean molecular mass near 500 km is not far from $M = 16$, i.e. corresponding to a major effect of atomic oxygen. An extrapolation up to altitudes higher than 1000 km with an atomic mass $M = 16$ cannot represent real atmospheric conditions. In fact, the mean molecular mass at 1000 km is always less than $M = 16$ and decreases to a value of the order of 4 when the thermopause temperature reaches 1000°K . This is due to the effect of helium which is a minor constituent in the lower thermosphere where $n(\text{He})/n(\text{N}_2) = 6.7 \times 10^{-6}$ but

TABLE XV - $\rho H^{1/2}$ (gm cm^{-5/2})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸	3.6x10 ⁻⁸
130	1.3	1.3	1.3	1.3	1.3	1.3	1.3	1.3	1.3
140	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.9x10 ⁻⁹	6.8x10 ⁻⁹
150	4.2	4.2	4.2	4.2	4.2	4.2	4.2	4.2	4.1
160	2.8	2.8	2.8	2.8	2.8	2.8	2.8	2.8	2.7
170	2.0	2.0	2.0	2.0	2.0	2.0	2.0	1.9	1.8
180	1.5	1.5	1.5	1.5	1.5	1.5	1.4	1.4	1.3
190	1.2	1.2	1.2	1.2	1.1	1.1	1.0	9.9x10 ⁻¹⁰	9.1x10 ⁻¹⁰
200	9.6x10 ⁻¹⁰	9.4x10 ⁻¹⁰	9.3x10 ⁻¹⁰	9.1x10 ⁻¹⁰	8.8x10 ⁻¹⁰	8.4x10 ⁻¹⁰	7.9x10 ⁻¹⁰	7.4	6.7
210	7.9	7.6	7.4	7.3	6.9	6.5	6.0	5.5	5.0
220	6.5	6.2	6.0	5.9	5.5	5.1	4.6	4.2	3.7
240	4.6	4.3	4.1	3.9	3.6	3.2	2.9	2.5	2.2
260	3.4	3.1	2.9	2.7	2.4	2.1	1.8	1.6	1.3
280	2.6	2.2	2.1	1.9	1.7	1.4	1.2	9.9x10 ⁻¹¹	8.0x10 ⁻¹¹
300	2.0	1.7	1.5	1.4	1.2	9.6x10 ⁻¹¹	7.9x10 ⁻¹¹	6.4	5.0
320	1.5	1.2	1.1	1.0	8.3x10 ⁻¹¹	6.7	5.3	4.2	3.2
340	1.2	9.6x10 ⁻¹¹	8.6x10 ⁻¹¹	7.6x10 ⁻¹¹	6.0	4.7	3.6	2.8	2.1
360	9.4x10 ⁻¹¹	7.4	6.5	5.7	4.4	3.4	2.5	1.9	1.4
380	7.5	5.8	5.0	4.4	3.3	2.4	1.8	1.3	9.6x10 ⁻¹²
400	6.0	4.5	3.8	3.3	2.4	1.8	1.3	9.2x10 ⁻¹²	6.5
420	4.9	3.6	3.0	2.6	1.8	1.3	9.2x10 ⁻¹²	6.5	4.5
440	4.0	2.8	2.4	2.0	1.4	9.7x10 ⁻¹²	6.7	4.6	3.2
460	3.2	2.3	1.9	1.6	1.1	7.3	4.9	3.3	2.2
480	2.7	1.8	1.5	1.2	8.2x10 ⁻¹²	5.5	3.6	2.4	1.6
500	2.2	1.5	1.2	9.9x10 ⁻¹²	6.5	4.2	2.8	1.8	1.2

TABLE XVI - Scale Height of Atomic Oxygen (km)

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	17.8	17.8	17.8	17.8	17.8	17.8	17.8	17.8	17.8
130	28.1	28.1	28.1	28.1	28.1	28.1	28.1	28.1	27.8
140	38.4	38.4	38.4	38.4	38.1	37.5	36.8	35.9	35.1
150	48.7	48.5	47.9	47.3	46.0	44.6	43.2	41.7	40.3
160	59.0	56.9	55.8	54.7	52.4	50.2	48.1	46.1	44.2
170	67.7	64.1	62.4	60.7	57.6	54.7	51.9	49.4	47.0
180	75.1	70.2	67.9	65.8	61.8	58.2	54.9	52.0	49.2
190	81.4	75.3	72.6	70.0	65.3	61.1	57.3	53.9	50.9
200	86.8	79.7	76.5	73.6	68.2	63.4	59.2	55.5	52.2
210	91.5	83.5	79.9	76.6	70.6	65.4	60.8	56.7	53.2
220	95.6	86.8	82.8	79.1	72.6	66.9	62.0	57.7	53.9
240	102.2	92.0	87.4	83.2	75.7	69.3	63.9	59.2	55.1
260	107.4	96.1	91.0	86.2	77.9	71.0	65.2	60.2	55.9
280	111.5	99.3	93.7	88.6	79.6	72.3	66.1	60.9	56.4
300	114.8	101.8	95.9	90.4	81.0	73.3	66.9	61.4	56.8
320	117.5	103.9	97.6	91.9	82.0	74.0	67.4	61.9	57.2
340	119.7	105.6	99.1	93.1	82.8	74.7	67.9	62.2	57.5
360	121.7	107.0	100.3	94.0	83.5	75.3	68.3	62.6	57.8
380	123.4	108.3	101.3	94.9	84.1	75.8	68.7	63.0	58.2
400	124.8	109.3	102.2	95.4	84.5	76.2	69.1	63.4	58.6
420	126.2	110.3	102.9	96.0	85.0	76.7	69.5	63.8	58.9
440	127.4	111.1	103.6	96.6	85.6	77.1	69.9	64.1	59.2
460	128.6	111.9	104.2	97.2	86.0	77.6	70.4	64.5	59.6
480	129.6	112.6	104.8	97.7	86.6	78.0	70.8	64.9	60.0
500	130.6	113.3	105.5	98.3	87.1	78.5	71.2	65.3	60.3

TABLE XVII - Scale Height of Molecular Nitrogen

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	10.2	10.2	10.2	10.2	10.2	10.2	10.2	10.2	10.2
130	16.0	16.0	16.0	16.0	16.0	16.0	16.0	16.0	15.9
140	21.9	21.9	21.9	21.9	21.8	21.4	21.0	20.5	20.0
150	27.8	27.7	27.4	27.0	26.3	25.5	24.7	23.8	23.0
160	33.7	32.5	31.9	31.2	30.0	28.7	27.5	26.3	25.2
170	38.7	36.6	35.7	34.7	32.9	31.2	29.7	28.2	26.9
180	42.9	40.1	38.8	37.6	35.3	33.3	31.4	29.7	28.1
190	46.5	43.0	41.5	40.0	37.3	34.9	32.8	30.8	29.1
200	49.6	45.6	43.7	42.0	39.0	36.2	33.8	31.7	29.8
210	52.3	47.7	45.7	43.8	40.4	37.4	34.7	32.4	30.4
220	54.6	49.6	47.3	45.2	41.5	38.2	35.4	33.0	30.8
240	58.4	52.6	50.0	47.6	43.2	39.6	36.5	33.8	31.5
260	61.4	54.9	52.0	49.3	44.5	40.6	37.2	34.4	31.9
280	63.7	56.7	53.6	50.6	45.5	41.3	37.8	34.8	32.2
300	65.6	58.2	54.8	51.7	46.2	41.9	38.2	35.1	32.5
320	67.1	59.4	55.8	52.5	46.8	42.3	38.5	35.4	32.7
340	68.4	60.3	56.6	53.2	47.3	42.7	38.8	35.6	32.9
360	69.5	61.1	57.3	53.7	47.7	43.0	39.0	35.8	33.0
380	70.5	61.9	57.9	54.2	48.0	43.3	39.2	36.0	33.3
400	71.3	62.4	58.4	54.5	48.3	43.6	39.5	36.2	33.5
420	72.1	63.0	58.8	54.9	48.6	43.8	39.7	36.4	33.7
440	72.8	63.5	59.2	55.2	48.9	44.1	40.0	36.6	33.8
460	73.5	63.9	59.5	55.5	49.2	44.3	40.2	36.8	34.0
480	74.1	64.3	59.9	55.8	49.5	44.6	40.4	37.1	34.3
500	74.6	64.7	60.3	56.1	49.8	44.9	40.7	37.3	34.4

TABLE XVIII - $n(O) H(O) \text{ cm}^{-2}$

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	1.35×10^{17}	1.35×10^{17}	1.35×10^{17}	1.35×10^{17}	1.35×10^{17}	1.35×10^{17}	1.35×10^{17}	1.35×10^{17}	1.32×10^{17}
130	8.71×10^{16}	8.71×10^{16}	8.71×10^{16}	8.71×10^{16}	8.74×10^{16}	8.74×10^{16}	8.74×10^{16}	8.72×10^{16}	8.51×10^{16}
140	6.45	6.45	6.46	6.46	6.47	6.46	6.44	6.39	6.20
150	5.13	5.14	5.14	5.12	5.12	5.08	5.02	4.95	4.77
160	4.28	4.26	4.24	4.22	4.19	4.12	4.04	3.96	3.77
170	3.66	3.62	3.60	3.56	3.50	3.42	3.32	3.22	3.04
180	3.19	3.13	3.09	3.05	2.97	2.87	2.76	2.65	2.48
190	2.82	2.74	2.69	2.64	2.54	2.44	2.32	2.20	2.03
200	2.51	2.41	2.36	2.30	2.20	2.08	1.96	1.84	1.68
210	2.25	2.14	2.08	2.02	1.91	1.79	1.66	1.54	1.39
220	2.03	1.91	1.85	1.78	1.66	1.54	1.42	1.30	1.16
240	1.67	1.54	1.47	1.40	1.28	1.16	1.04	9.28×10^{15}	8.08×10^{15}
260	1.39	1.25	1.18	1.11	9.91×10^{15}	8.74×10^{15}	7.66×10^{15}	6.68	5.67
280	1.16	1.02	9.57×10^{15}	8.92×10^{15}	7.74	6.66	5.68	4.83	4.00
300	9.83×10^{15}	8.44×10^{15}	7.79	7.18	6.07	5.08	4.23	3.50	2.82
320	8.32	6.99	6.38	5.80	4.78	3.90	3.16	2.55	2.00
340	7.08	5.81	5.24	4.70	3.77	3.00	2.37	1.86	1.42
360	6.04	4.84	4.31	3.82	2.98	2.31	1.77	1.36	1.01
380	5.16	4.04	3.55	3.11	2.36	1.78	1.33	9.91×10^{14}	7.19×10^{14}
400	4.41	3.38	2.94	2.53	1.87	1.38	1.00	7.27	5.14
420	3.79	2.84	2.43	2.06	1.49	1.07	7.56×10^{14}	5.34	3.68
440	3.25	2.38	2.01	1.69	1.18	8.27×10^{14}	5.71	3.93	2.63
460	2.80	2.00	1.67	1.38	9.43×10^{14}	6.43	4.32	2.89	1.89
480	2.40	1.69	1.39	1.13	7.53	5.00	3.27	2.14	1.36
500	2.07	1.42	1.16	9.29×10^{14}	6.01	3.90	2.48	1.58	9.82×10^{13}

TABLE XIX - $n(N_2) H(N_2) \text{ cm}^{-2}$

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	5.92×10^{17}	5.93×10^{17}	5.93×10^{17}	5.93×10^{17}	5.93×10^{17}	5.93×10^{17}	5.93×10^{17}	5.93×10^{17}	5.79×10^{17}
130	2.74	2.74	2.74	2.74	2.75	2.75	2.75	2.74	2.67
140	1.62	1.62	1.62	1.62	1.62	1.62	1.61	1.59	1.53
150	1.08	1.08	1.08	1.08	1.07	1.06	1.04	1.01	9.65×10^{16}
160	7.83×10^{16}	7.78×10^{16}	7.73×10^{16}	7.66×10^{16}	7.54×10^{16}	7.34×10^{16}	7.10×10^{16}	6.82×10^{16}	6.40
170	5.96	5.84	5.76	5.67	5.50	5.28	5.02	4.74	4.37
180	4.68	4.51	4.42	4.32	4.12	3.88	3.63	3.37	3.05
190	3.75	3.56	3.46	3.34	3.13	2.90	2.66	2.43	2.15
200	3.06	2.85	2.74	2.63	2.42	2.20	1.98	1.77	1.54
210	2.52	2.30	2.20	2.09	1.89	1.68	1.48	1.30	1.11
220	2.10	1.88	1.78	1.67	1.48	1.29	1.12	9.60×10^{15}	8.00×10^{15}
240	1.48	1.28	1.19	1.09	9.29×10^{15}	7.79×10^{15}	6.46×10^{15}	5.31	4.24
260	1.07×10^{15}	8.87×10^{15}	8.06×10^{15}	7.28×10^{15}	5.93	4.76	3.78	2.97	2.27
280	7.82×10^{15}	6.24	5.55	4.91	3.83	2.94	2.23	1.68	1.22
300	5.78	4.43	3.86	3.34	2.49	1.83	1.33	9.51×10^{14}	6.64×10^{14}
320	4.30	3.17	2.70	2.29	1.63	1.14	7.92×10^{14}	5.43	3.61
340	3.22	2.28	1.90	1.58	1.07	7.19×10^{14}	4.75	3.11	1.97
360	2.43	1.65	1.35	1.09	7.08×10^{14}	4.53	2.86	1.78	1.08
380	1.84	1.20	9.59×10^{14}	7.58×10^{14}	4.69	2.87	1.72	1.03×10^{13}	5.96×10^{13}
400	1.39	8.76×10^{14}	6.84	5.27	3.11	1.82	1.04	5.94×10^{13}	3.29
420	1.06	6.41	4.89	3.68	2.07	1.16	6.34×10^{13}	3.45	1.82
440	8.07×10^{14}	4.70	3.50	2.58	1.38	7.39×10^{13}	3.86	2.00	1.01
460	6.18	3.45	2.52	1.81	9.25×10^{13}	4.73	2.36	1.17	5.66×10^{12}
480	4.73	2.54	1.81	1.27	6.21	3.03	1.44	6.85×10^{12}	3.17
500	3.62	1.88	1.31	8.94×10^{13}	4.17	1.95	8.87×10^{12}	4.02	1.78

TABLE XX - $n(O_2) H(O_2) \text{ cm}^{-2}$

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
120	1.05×10^{17}	1.05×10^{17}	1.05×10^{17}	1.05×10^{17}	1.05×10^{17}	1.05×10^{17}	1.05×10^{17}	1.05×10^{17}	1.03×10^{17}
130	4.36×10^{16}	4.36×10^{16}	4.36×10^{16}	4.36×10^{16}	4.38×10^{16}	4.39×10^{16}	4.38×10^{16}	4.36×10^{16}	4.25×10^{16}
140	2.38	2.38	2.39	2.39	2.40	2.39	2.37	2.34	2.25
150	1.51	1.51	1.50	1.50	1.49	1.47	1.44	1.40	1.33
160	1.04	1.03	1.03	1.02	9.97×10^{15}	9.68×10^{15}	9.31×10^{15}	8.90×10^{15}	8.29×10^{15}
170	7.62×10^{15}	7.44×10^{15}	7.33×10^{15}	7.20×10^{15}	6.95	6.63	6.26	5.87	5.36
180	5.77	5.54	5.41	5.26	4.99	4.66	4.32	3.97	3.55
190	4.48	4.22	4.08	3.93	3.65	3.34	3.04	2.73	2.39
200	3.55	3.27	3.13	2.99	2.71	2.43	2.16	1.90	1.62
210	2.84	2.57	2.43	2.30	2.04	1.79	1.55	1.33	1.11
220	2.30	2.04	1.91	1.78	1.55	1.33	1.12	9.43×10^{14}	7.69×10^{14}
240	1.55	1.31	1.20	1.10	9.08×10^{14}	7.42×10^{14}	6.00×10^{14}	4.79	3.72
260	1.06×10^{14}	8.60×10^{14}	7.71×10^{14}	6.86×10^{14}	5.43	4.22	3.24	2.46	1.82
280	7.45×10^{14}	5.75	5.03	4.37	3.29	2.43	1.77	1.28	8.97×10^{13}
300	5.27	3.88	3.32	2.63	2.01	1.41	9.78×10^{13}	6.70×10^{13}	4.45
320	3.75	2.65	2.21	1.82	1.24	8.25×10^{13}	5.42	3.52	2.22
340	2.70	1.82	1.48	1.19	7.66×10^{13}	4.85	3.02	1.86	1.11
360	1.95	1.26	9.96×10^{13}	7.81×10^{13}	4.76	2.86	1.69×10^{12}	9.85×10^{12}	5.58×10^{12}
380	1.41	8.71×10^{13}	6.73	5.14	2.97	1.69	9.46×10^{12}	5.24	2.82
400	1.03	6.06	4.57	3.39	1.86	1.00	5.33	2.80	1.43
420	7.55×10^{13}	4.24	3.11	2.25	1.16	5.99×10^{12}	3.01	1.50	7.28×10^{11}
440	5.51	2.97	2.12	1.49	7.34×10^{12}	3.58	1.71	8.08×10^{11}	3.72
460	4.06	2.09	1.46	9.91×10^{12}	4.63	2.14	9.71×10^{11}	4.36	1.91
480	2.99	1.47	9.98×10^{12}	6.65	2.93	1.29	5.54	2.37	9.83×10^{10}
500	2.20	1.03	6.87	4.45	1.86	7.79×10^{11}	3.17	1.29	5.08

becomes a principal constituent above 750 km since $n(\text{He})/n(\text{N}_2) > 1$ by diffusion. Since its effect increases when the temperature decreases, helium must be introduced in any atmospheric model above 500 km. In other words, the heterosphere above 500 km cannot be studied without introducing the effect of helium.

A detailed analysis of the behavior of helium requires the study of its escape from the terrestrial atmosphere [see, for example, Nicolet (1957), Bates and McDowell (1957) and references infra]. If, for example, we consider the life-time of helium (time necessary to reduce its concentration to 50 percent at the base of the isothermal layer) as the escape parameter the following values are obtained:

Scale Height (km)	Temperature ($^{\circ}\text{K}$)	Life-time (years)
400	1700	13
420	1790	6.5
440	1875	3.3
460	1960	1.8
480	2050	1.1
500	2130	0.65

Such life-times show that the temperature is an important parameter, since it varies during a solar cycle. However, for a simple computation of an atmospheric model the essential fact must be that diffusion supports the flow needed to maintain a diffusion equilibrium. In other words, it is necessary

to know the conditions when a source is involved (Mange 1961). A computation made by Kockarts (1962) for the conditions defined in Table IV shows that maximum flow reaching about 10^8 atoms $\text{cm}^{-2}\text{sec}^{-1}$ is allowed. Since a source of 10^6 $\text{cm}^{-2}\text{sec}^{-1}$ must represent a maximum, the vertical distribution of helium in an atmospheric model is given by the diffusion equilibrium.

Another important factor is the "diffusion level" which corresponds, in a computation, to the beginning of diffusion. The time of diffusion given by Mange's formula (26) shows that a difference of altitude of 5 km in the "diffusion level" leads to a difference of a factor of the order of 2 in the times of diffusion which correspond to a ratio of 2 for the concentrations of helium in the whole diffusion region. There is no disagreement with the observational evidence if an altitude corresponding to (110 ± 5) km for the diffusion level is adopted. The author, in collaboration with Kockarts, has considered various possibilities. A detailed analysis cannot be given here and the conditions which correspond to a diffusion level of 105 km are adopted for the physical conditions of the heterosphere between 500 km and 3000 km.

The problem of atomic hydrogen must be considered in the same way as that of helium, but the solution which is obtained is different (Mange, 1955, 1961) (Bates and Patterson, 1961 a and b). The equilibrium distribution of atomic hydrogen is subject to distribution by flow and leads to a scale height of atomic hydrogen in the main atmosphere which

is far smaller than the scale height of hydrogen in diffusion equilibrium. A detailed analysis has been made by Kockarts (1962) in using the atmospheric conditions defined by Table VI. If the total content in atomic hydrogen above 200 km corresponds to about 1×10^{12} , 3×10^{12} and 6×10^{12} atoms cm^{-2} it is possible to consider the level of diffusion at 115, 110 and 105 km; respectively. The first value corresponds to the initial determination from the value of the profile of solar Lyman- α by Purcell and Tousey (1961). The second value is somewhat more than the value deduced by Bates and Patterson (1961b); namely $1.8 \times 10^{12} \text{ cm}^{-2}$. The third value (see Johnson and Fisk, 1960) can be considered as one which leads to a maximum value for the atomic hydrogen concentration (Thomas and Donahue, 1961) in the region where helium plays its essential role. In any case, the total content of hydrogen atoms under these circumstances is not less than $5 \times 10^{12} \text{ cm}^{-2}$ above 100 km, corresponding to a concentration at 100 km not far from 10^7 cm^{-3} . Such a value of the atomic hydrogen concentration at 100 km may be considered as a round figure corresponding to a conservative value ($\pm 50\%$) for the mixing region. In any case, we adopt here the conditions which lead to the maximum concentration of atomic hydrogen, since our purpose is to show that, in any circumstance, there is a helium layer between the regions where atomic oxygen and atomic hydrogen are the principal constituents.

According to these conditions, the helium concentration at an altitude of 500 km is $(2 \pm 0.3) \times 10^6 \text{ cm}^{-3}$. With such

concentrations, atomic hydrogen cannot have, below 2000 km, a detectable effect in an analysis of the atmospheric density. Table XXI gives the density in gm cm^{-3} from 500 km to 3000 km. The small difference at 500 km between the density values of Table XXI and Table VI corresponds to the helium effect increasing with altitude. The presence of hydrogen, indicated by the values of the second line, has only a practical effect on the value of the density for sufficiently low temperatures above 2000 km.

It must be pointed out that, for temperatures greater than 1400°K , the density at 1500 km is greater than $1 \times 10^{-18} \text{ gm cm}^{-3}$, i.e. decreases only by a factor of the order of 1000 between 500 km and 1500 km. At 3000 km the density is greater than $10^{-19} \text{ gm cm}^{-3}$ for $T \geq 1500^\circ\text{K}$. Even for a temperature of the order of 1000°K , the density at 3000 km is not less than $10^{-20} \text{ gm cm}^{-3}$.

The total concentration between 500 km and 3000 km is given in Table XXII. When two values are given for a certain altitude they correspond to the total concentrations without and with atomic hydrogen; respectively. For $T \geq 1500^\circ\text{K}$, the total concentration at 1000 km is not less than 10^6 cm^{-3} and at 3000 km still reaches 10^4 cm^{-3} .

The effect of each constituent is shown in Table XXIII giving the mean molecular mass from 500 km to 3000 km. The effect of molecular nitrogen still exists above 500 km when the temperature is sufficiently high ($M > 16$). The presence of helium is certain at 500 km when the temperature is low

TABLE XXI - Density (gm cm^{-3}) between 500 km and 3000 km

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
500*	6.66×10^{-15}	4.79×10^{-15}	4.00×10^{-15}	3.31×10^{-15}	2.26×10^{-15}	1.54×10^{-15}	1.05×10^{-15}	7.07×10^{-16}	4.69×10^{-16}
520	5.45	3.85	3.19	2.62	1.75	1.17	7.76×10^{-16}	5.15	3.35
540	4.50	3.12	2.56	2.08	1.36	8.93×10^{-16}	5.80	3.78	2.41
560	3.72	2.54	2.06	1.66	1.06	6.85	4.36	2.78	1.74
580	3.07	2.08	1.67	1.33	8.34×10^{-16}	5.28	3.29	2.06	1.27
600	2.56	1.72	1.36	1.07	6.58	4.09	2.50	1.54	9.30×10^{-17}
650	1.66	1.07	8.25×10^{-16}	6.32×10^{-16}	3.70	2.19	1.28	7.55×10^{-17}	4.42
700	1.11	6.79×10^{-16}	5.12	3.82	2.13	1.21	6.76×10^{-17}	3.86	2.22
750	7.49×10^{-16}	4.39	3.22	2.35	1.25	6.82×10^{-17}	3.69	2.07	1.20
800	5.14	2.88	2.07	1.47	7.49×10^{-17}	3.95	2.10	1.18	7.02×10^{-18}
850	3.58	1.91	1.34	9.37×10^{-17}	4.59	2.37	1.26	7.20×10^{-18}	4.49
900	2.52	1.29	8.86×10^{-17}	6.06	2.88	1.47	7.93×10^{-18}	4.73	3.11
1000	1.28	6.08×10^{-17}	4.03	2.68	1.24	6.51×10^{-18}	3.78	2.48	1.78
1250	2.82×10^{-17}	1.19	7.73×10^{-18}	5.24×10^{-18}	2.79×10^{-18}	1.79	1.24	9.04×10^{-19}	6.71×10^{-19}
1500	8.20×10^{-18}	3.68×10^{-18}	2.59	1.94	2.82	1.83	1.28	9.42	7.13
2000	1.79	1.05	8.14×10^{-19}	6.49×10^{-19}	1.22	8.48×10^{-19}	5.93×10^{-19}	4.17	2.94
	1.81**	1.07	8.40	6.75	1.25	8.80	6.25	4.48	3.28
2500	8.16×10^{-19}	4.92×10^{-19}	3.73	2.86	4.10×10^{-19}	2.64	1.67	1.05	6.65×10^{-20}
	8.36	5.13	3.94	3.06	4.33	2.89	1.91	1.28	8.97
3000	4.61	2.61	1.90	1.39	1.64	9.60×10^{-20}	5.48×10^{-20}	3.12×10^{-20}	1.79
	4.78	2.79	2.07	1.56	1.82	1.15×10^{-19}	7.23	4.75	3.45×10^{-21}
					7.24×10^{-20}	3.89×10^{-20}	2.02	1.05	5.49×10^{-21}
					8.74	5.37	3.39	2.30	1.79×10^{-20}

(*) Difference with Table VII, effect of helium.

(**) Second line with effect of atomic hydrogen.

TABLE XXII - Total concentration from 500 km to 3000 km (cm^{-3})

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
500*	2.1×10^8	1.6×10^8	1.3×10^8	1.1×10^7	8.0×10^7	5.6×10^7	3.9×10^7	2.7×10^7	1.9×10^7
520	1.8	1.3	1.1	9.1×10^7	6.3	4.3	3.0	2.0	1.4
540	1.5	1.1	8.9×10^7	7.3	5.0	3.4	2.2	1.5	1.0
560	1.2	8.8×10^7	7.3	5.9	3.9	2.6	1.7	1.1	7.4×10^6
580	1.0	7.3	6.0	4.8	3.1	2.0	1.3	8.6×10^6	5.6
600	8.8×10^7	6.1	4.9	3.9	2.5	1.6	1.0	6.6	4.3
650	5.9	3.9	3.1	2.4	1.5	9.1×10^6	5.6×10^6	3.6	2.3
700	4.0	2.6	2.0	1.3	8.8×10^6	5.3	3.2	2.1	1.4
750	2.8	1.7	1.3	9.6×10^6	3.4	3.2	2.0	1.3	9.1×10^5
800	2.0	1.1	8.5×10^6	6.3	3.5	2.1	1.3	8.8×10^5	6.4
850	1.4	7.9×10^6	5.7	4.2	2.3	1.4	9.1×10^5	6.4	4.8
900	1.0	5.5	4.0	2.9	1.6	1.0	6.8	4.9	3.8
1000	5.4×10^6	2.9	2.0	1.5	8.8×10^5	5.9×10^5	4.2	3.2	2.5
1250	1.5	8.3×10^5	6.2×10^5	4.9×10^5	9.1	6.1	4.5	3.4	2.8
1500	6.3×10^5	3.8	6.4	5.1	3.3	2.4	1.8	1.3	1.0
2000	2.3	4.0	3.2	2.5	3.5	2.6	2.0	1.6	1.3
2500	1.1	1.5	1.2	2.7	1.9	1.4	8.9×10^4	6.2×10^4	4.4×10^4
3000	1.3**	8.6	6.8	9.7×10^4	6.1×10^4	4.0×10^4	1.1×10^4	8.1	6.4
	6.9×10^4	3.9×10^4	2.9×10^4	1.2×10^5	7.5	5.4	2.5×10^4	1.6	1.0
	7.9	5.0	3.9	4.3×10^4	2.5	1.4	3.9	2.9	2.4
				4.3×10^4	3.5	2.5	8.2×10^3	4.7×10^3	2.7×10^3
				5.5	1.1	5.9	1.9×10^3	1.4×10^3	1.3×10^2
				2.1	2.0	1.5×10^4	3.0×10^3	1.6×10^3	8.3×10^2
				3.1			1.1×10^4	9.0×10^3	8.2×10^3

(*) Difference with Table IX, effect of helium.

(**) Second line with effect of atomic hydrogen.

TABLE XXIII - Mean Molecular Mass between 500 km and 3000 km (Atomic oxygen = 16)

Altitude -(km)	6D	1	1.5	2	3	4	5	6	7
500*	18.9	18.2	17.9	17.6	17.0	16.6	16.1	15.7	15.3
520	18.6	17.9	17.6	17.3	16.8	16.3	15.8	15.4	14.9
540	18.3	17.6	17.4	17.0	16.5	16.0	15.6	15.1	14.5
560	18.0	17.4	17.1	16.8	16.3	15.8	15.3	14.8	14.1
580	17.8	17.2	16.9	16.6	16.0	15.5	15.0	14.4	13.6
600	17.6	17.0	16.7	16.4	15.8	15.3	14.7	13.9	13.0
650	17.1	16.5	16.2	15.9	15.2	14.6	13.7	12.7	11.4
700	16.6	16.0	15.7	15.4	14.6	13.7	12.5	11.2	9.6
750	16.2	15.6	15.2	14.8	13.8	12.6	11.1	9.6	8.0
800	15.8	15.1	14.7	14.2	12.9	11.4	9.7	8.0	6.6
						11.2	9.5	7.8	6.3
850	15.5	14.6	14.1	13.4	11.9	10.1	8.3	6.8	5.6
						9.9	8.0	6.5	5.3
900	15.1	14.1	13.4	12.6	10.7	8.8	7.1	5.8	5.0
						8.6	6.8	5.5	4.7
1000	14.2	12.7	11.8	10.7	8.5	6.7	5.4	4.7	4.4
				10.5	8.3	6.4	5.1	4.4	4.0
1250	11.2	8.7	7.5	6.4	5.1	4.4	4.2	4.1	4.0
	11.0**	8.5	7.3	6.2	4.8	4.2	3.8	3.6	3.4
1500	7.9	5.8	5.1	4.6	4.2	4.1	4.0	4.0	4.0
	7.7	5.6	4.9	4.4	3.9	3.7	3.5	3.3	3.1
2000	4.7	4.2	4.1	4.0	4.0	4.0	4.0	4.0	4.0
	4.5	3.9	3.8	3.6	3.5	3.2	2.9	2.6	2.3
2500	4.1	4.0	4.0	4.0	4.0	4.0	4.0	4.0	4.0
	3.8	3.6	3.5	3.3	3.1	2.7	2.3	2.0	1.6
3000	4.0	4.0	4.0	4.0	4.0	4.0	4.0	4.0	4.0
	3.6	3.4	3.2	3.0	2.7	2.2	1.8	1.5	1.3

(*) Difference with Table VIII, effect of helium.

(**) Second line with effect of atomic hydrogen.

($M < 16$). Atomic oxygen has an effect for $M > 4$. The introduction of hydrogen is shown by the values of the second line for various altitudes. Its effect is apparent at 2000 km only when the temperature is sufficiently low. The range of variations from $M = 19$ to $M = 1.3$ show how difficult it is to detect the exact composition of the heterosphere above 500 km if only density data are used. Finally Table XXIV gives the atmospheric scale height between 500 km and 3000 km corresponding to the simultaneous effect of all constituents (N_2 , O, He and H). The values at 500 km are essentially the values associated with atomic oxygen. At 750 km, $H = (125 \pm 10)$ km is remarkably constant due to the small variation of T/M ; corresponding to the effect of helium. The two values which represent the atmosphere without and with atomic hydrogen show how the scale height increases when the ratio T/M increases at low temperatures.

In conclusion, the various Tables giving all parameters for a neutral atmosphere correspond to a consistent atmospheric model. They do not represent all possible conditions, but must be considered as average conditions. Thus, the picture of the general behavior of the heterosphere is obtained. Other calculations, as yet unpublished, computed in collaboration with Mr. Kockarts and Mr. Dogniaux, will show how other boundary conditions can change the conclusions or can refine the problem.

TABLE XXIV - Atmospheric scale height (km)

Altitude (km)	6D	1	1.5	2	3	4	5	6	7
500*	110.8	99.5	94.4	89.7	81.8	75.9	70.6	66.4	63.2
520	113.4	101.6	96.4	91.6	83.6	77.6	72.3	68.1	65.0
540	115.9	103.9	98.4	93.6	85.4	79.2	74.0	70.0	67.2
560	118.3	105.9	100.4	95.4	87.1	80.9	75.8	72.0	69.7
580	120.7	108.0	102.3	97.3	88.8	82.7	77.8	74.3	72.8
600	123.1	109.9	104.2	99.2	90.6	84.6	80.0	77.1	76.2
650	128.8	114.8	109.0	103.8	95.5	90.1	86.8	86.0	88.3
700	134.0	119.8	113.8	108.7	95.7	90.4	87.2	86.8	89.7
750	139.3	124.9	119.1	109.0	101.1	97.3	96.3	98.9	106.1
800	144.6	125.1	119.4	114.4	108.3	106.8	109.9	100.4	108.8
850	144.8**	130.6	125.1	114.7	108.8	107.8	111.5	117.4	130.3
900	150.2	130.8	125.5	121.2	117.5	119.9	128.4	120.1	135.2
1000	150.5	136.9	132.3	121.7	118.4	121.5	131.1	141.7	159.2
1250	171.4	164.9	166.4	169.6	189.2	163.5	187.3	211.4	232.0
1500	232.1	256.9	277.8	302.7	338.7	347.4	335.5	315.6	294.2
2000	351.7	412.0	434.5	446.3	434.6	371.8	367.0	354.4	345.8
2500	665.7	641.6	611.5	578.7	515.6	404.7	371.4	341.8	316.2
3000	914.1	843.4	813.1	788.5	754.2	776.0	817.7	880.8	978.4
	972.8	842.8	784.3	733.1	647.8	583.9	530.0	486.0	448.7
	1077	1005	983.0	974.9	977.3	1063.0	1166	1269	1372

(*)Difference with Table XI, effect of helium.

(**)Second line with effect of atomic hydrogen.

12. Conclusions

When a comparison between the densities given in Tables VII and XXI and published data on densities deduced from satellite air-drag (Jacchia 1960, 1961 a, King-Hele and Walker 1961, Martin, Neveling, Priester and Roemer 1961, Priester 1961) is made, it is clear that the absolute value of the density is related to the thermopause temperature (Nicolet 1961 d). The general trend of the diurnal variation of the density is shown by the variation of temperature in an atmosphere subject primarily to a vertical diffusion of the various gases and to a vertical conduction of heat. Therefore, the atmospheric scale height being a complicated function of T/M and its variable gradient, having several maxima, can be interpreted in a simple way without any additional hypotheses. In the same way, the variation of the density associated with that of solar activity is a temperature effect. The correlation observed between the 10.7 cm solar flux measured at Ottawa and fluctuations in the orbital period of various satellites (Priester 1959, Jacchia 1959 a) is a clear indication of the variation of the ultraviolet radiation leading to variation of the temperature in the heterosphere (Nicolet, 1960 c). The correlation found by Jacchia (1959 b, 1961 a) between the worldwide variation of density and magnetic storms is also an indication of a general increase of the temperature which must be explained by a non corpuscular process as well as the hydro-magnetic heating introduced by Dessler (1959).

Thus, the temperature is the essential parameter to be

considered when diffusion and conduction have been introduced before looking for special processes to explain the behavior of the heterosphere. A recent analysis made by Jacchia (1961 b) now gives a clear picture of the principal effects which can be deduced from the variations of the satellite accelerations. From satellites with perigees between 350 km and 660 km, Jacchia found that the night-time temperature T_{mn} between 1958 and 1961 is related to the solar flux S [unit, 10^{-22} watt m^{-2} (cycle sec^{-1}) at 10.7 cm] in a first approximation as follows:

$$T_{mn} = (555 + 3S)^{\circ}K \quad (32)$$

When the solar flux decreased from $S = 350$ to $S = 90$, the night-time temperature of the thermopause decreased by about $800^{\circ}K$, i.e. varied from $1600^{\circ}K$ to $800^{\circ}K$. The relation between the temperature at the center of the diurnal bulge, T_{MD} , and the night-time temperature T_{mn}

$$T_{MD}/T_{mn} = 1.35 \quad (33)$$

shows a difference between day and night of between $300^{\circ}K$ and $500^{\circ}K$.

Furthermore, Jacchia (1961 b), considering the relationship between the density $\rho(z)$ and the solar flux $S^m(z)$, deduced values of $m(z)$ which correspond to the values obtained from observational data (Priester, 1961). Finally, Jacchia (1961 b) showed that the effect of magnetic storms can be represented by a simple equation, in a first approximation,

$$\Delta T = (2.5 a_p)^{\circ}K \quad (34)$$

where a_p denotes the geomagnetic planetary index.

Thus, the atmospheric model with its various parameters given in Tables VI to XXIV can provide a preliminary basis to more adequately study atmospheric phenomena. Various observations, such as the direct determination of the temperature (Blamont et al. 1961) or the electron temperatures (Serbu, Bourdeau and Donley, 1961), could lead to an exact calibration when the total density is known. The determination of the electronic or ionic composition (Gringauz 1961, Istomin 1961, Hanson 1962, Bourdeau et al. 1962, Bauer and Jackson 1962) is another aspect which can be introduced in order to determine the complete behavior of the heterosphere.

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